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Boris S. Sokolov

Academician, Russian Academy of Sciences
Counsellor of the Presidium of the Russian Academy of Sciences

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Contents

Vol. 3, No. 5, 1995

Simultaneous English language translation of the journal is available from МАИК Наука/Interperiodica Publishing (Russia). *Stratigraphy and Geological Correlation* ISSN 0869-5938.

Dagmara Maksimilianovna Rauser-Chernousova, Congratulations on Her 100th Birthday 427

The Scientific and Organizational Activities of D. M. Rauser-Chernousova

B. V. Poyarkov 428

Micropaleontological Research of the Mesozoic and Cenozoic at the Geological Institute, Russian Academy of Sciences

K. I. Kuznetsova and N. G. Muzylev 432

Facial-Ecological Structure of the Derevnya Formation Microbiota: Upper Riphean, Turukhansk Uplift, Siberia

P. Yu. Petrov and A. F. Veis 435

Carbon and Oxygen Isotopic Variations in the Lower Proterozoic Carbonates of the Kola Peninsula

B. G. Pokrovskii and V. A. Melezhik 461

Isotopic Datings of the Lower Cambrian Clays from Estonia and Their Interpretation

I. M. Gorokhov, E. P. Kutyavin, N. N. Mel'nikov, and T. L. Turchenko 473

Regional Middle-Upper Cambrian Stratigraphic Scheme of the East European Platform

N. A. Volkova and V. V. Kir'yanov 484

Paleozoic Deposits in Kolguev Island (Barents Sea)

E. N. Preobrazhenskaya, V. I. Ustritskii, and E. G. Bro 493

Late Jurassic Benthic Assemblages of Siberia

V. A. Zakharov 504

Subdivision of Continental Neogene Deposits of the Far East Based on Diatoms

A. I. Moiseeva 510

Correlation of Loess-Soil Sections in Northern Bulgaria from Magnetic and Paleomagnetic Data

V. A. Bol'shakov and A. A. Svitoch 521

Stratigraphy and Geological Correlation, Vol. 3, No. 5, 1995, p. 427. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 3, No. 5, 1995, p. 3.
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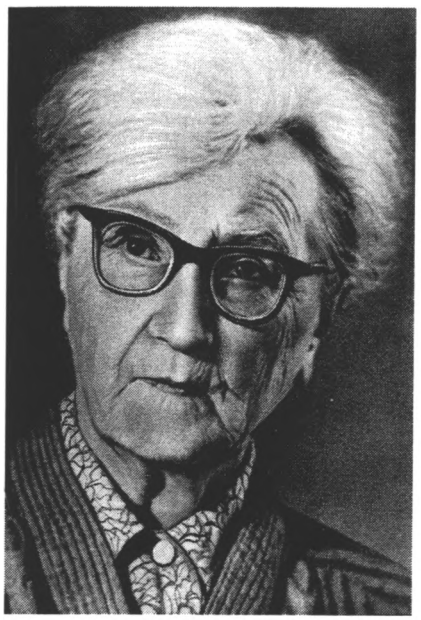
Dagmara Maksimilianovna Rauser-Chernousova, Congratulations on Her 100th Birthday



March 19, 1995, was the 100th birthday of Dagmara Maksimilianovna Rauser-Chernousova, a scientist known the world over and the initiator of micropaleontological studies in Russia. She was one of the first to realize that microfauna hold great significance for Paleozoic stratigraphy, and she was the first to prove it.

Practice is the criterion of theory, the order of the Combat Red Banner, an award with which she was honored for elaborating the Carboniferous-Permian stratigraphy of the Ishimbai petroleum field in Bashkiria during World War II, is excellent proof of this statement. Her zonation schemes of the Upper Paleozoic in the Volga–Urals region are still in practical use.

Rauser-Chernousova is a famous theorist of the paleobiology, taxonomy, and morphology of Protozoa. She is an amazing pedagogue, and many micropaleontologists in Russia are proud to be her students. Using her excellent organizing skills, Rauser-Chernousova founded the Micropaleontological Laboratory at the Geological Institute of the Russian Academy of Sciences, where she retained her leadership for a very long time. She also organized the All-Union Micropaleontological Commission, and was an unsurpassed editor for the majority of issues of the former journal *Voprosy Mikropaleontologii.*

Rauser-Chernousova was awarded the Lenin, Karpinsky, and Cushman Foundation prizes. She is also an Honored Scientist, a Professor, a Doctor of the geosciences, and an Honored Professor of the International Science Foundation.

Even more important and impressive, however, is that, throughout her entire life, Rauser-Chernousova showed unselfish devotion to the sciences, standing as a symbol of human probity and kindness. We wish to you, dear Dagmara Maksimilianovna, many happy years ahead.

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The Scientific and Organizational Activities of D. M. Rauser-Chernousova

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Abstract – Rauser-Chernousova's ideas on the study of Paleozoic foraminifers and the key trends in her scientific and organizational activities are discussed. Rauser-Chernousova's scientific works as well as the impressions that she has made on the author over 40 years are used as the basis for the present paper.
Key words: **Paleozoic foraminifers, micropaleontology.**

A jubilee meeting of the Scientific Council of the Geological Institute (GIN), Russian Academy of Sciences, was held on March 21, 1995, to celebrate the 100th birthday (March 19, 1995) of prominent micropaleontologist Professor D.M. Rauser-Chernousova, winner of the Lenin Prize and Doctor of Geology. I can not recall a similar agenda in the 50-year history of the GIN Scientific Council, and it is rare that a research institute can celebrate the 100th birthday of an active researcher.

Colleagues have appraised Rauser-Chernousova's creative labour at its true worth: she was awarded the Karpinsky Medal by the Russian Academy of Sciences and the Cushman Medal by the American Geological Society. Rauser-Chernousova is the only scientist who has been honored with the two highest awards in paleontology.

When future historians of science begin to study the 20th century, they will be surprised by the blossoming of ideas in micropaleontology that happened in a country spanning one-sixth of all continents (the former USSR) and a time period spanning two-thirds of a century. In studying this phenomenon, they should first trace its implications for geology. It is well-known that geology, as the oldest field of knowledge, became a science after it began to consider geologic events in time and space. Introducing the time scale into geology as a main research tool became a prerogative of paleontology for a long period of time. Micropaleontology widened the limits of the paleontological method in stratigraphy just by the mass character of the actual material. Due to their small dimensions, these materials are easily obtainable even from drilling cores.

Just as it is difficult to follow a single stream in a mountain river flow, it is not an easy matter to single out one discovery among the other important ones made by Rauser-Chernousova, because the main feature of her work, which covers two-thirds of this century, is that it all integrated. I would like to present some of my personal impressions about certain aspects of her scientific and organizational activities.

I heard the name Rauser-Chernousova for the first time in 1950 from A.D. Miklukho-Maklai while in the Guzan Mountains, situated in southern Fergana. As a 20-year-old student from a provincial institute, I did not realize at that time, that 45 years later, I would be speaking about this outstanding researcher in front of those gathered in the GIN conference hall. Later, in 1955, I was introduced by Miklukho-Maklai to Rauser-Chernousova as his post-graduate student during a break between sessions of the All-Union Paleontological Society (AUPS). Five years later, Rauser-Chernousova invited me to take part in All-Union Micropaleontological meetings, and I began writing articles for the journal *Voprosy Mikropaleontologii*. In the winter of 1969, I visited her country house for a review of my doctorate thesis. When I entered the room, I saw my work on the table and a thick pile of sheets near it with notes made in tiny handwriting with different color inks. Seeing the surprise in my eyes, Rauser-Chernousova smiled and said, "First of all, the work basically turned out well and I liked it. As for my notes, I have written them in different colors to avoid reading the manuscript twice. One color is for remarks, the second color marks items that I should mention in my review, and the third color marks what may appear useful in my future work." Then, every proposition of my thesis was discussed very thoroughly, in a gentle manner, without her own opinion being forced on me, and without any offence being taken. After defending my thesis, we met again a few years later in 1975 at a session of the Carboniferous Congress held at Moscow State University. Again, our conversation helped me to find my way with my new status. Over 40 years we met many times.

Rauser-Chernousova is striking in her rare and surprising combination of intuition together with the systematic character and consistency in her work. In addition to Rauser-Chernousova, Paleozoic foraminifers

were studied in the 1930s by G.A. Dutkevich, M.A. Kalmykova, L.P. Grozdilova, A.D. Miklukho-Maklai, and other scientists. Each of them contributed greatly to the research of foraminifers, but Rauser-Chernousova achieved a more profound success. I guess her intuition served its purpose.

At the very beginning of her career, Rauser-Chernousova recognized the importance of this fauna for stratigraphy; she also realized the potential inherent in studying foraminifers and in cooperation between researchers' efforts. She tirelessly directed herself not only to increasing the staff of the laboratory she was in charge of, but also to organizing a micropaleontological group in every oil survey and prospecting institution. As a result, the number of such groups grew to 160 - 170, with more than 200 working micropaleontologists. The support of such prominent geologists as academicians A.D. Arkhangel'skii, N.S. Shatskii and V.V. Menner contributed greatly to this. They recognized the importance of paleontology and stratigraphy for the advancement of geology and the creation of a raw material base; they also highly appraised Rauser-Chernousova's manner and methods of her collective work.

The joint efforts of scientists and those who worked at production enterprises eventually promoted the blossoming of micropaleontology. Rauser-Chernousova put this approach for organizing scientific research into practice, which is evidenced by the 30 volumes of *Voprosy Mikropaleontologii* issued during the 40 years that she worked as an editor.

Rauser-Chernousova contributed much to the training of specialists, providing them with the necessary literature for determining foraminifers she also undertook the responsibility of improving the experience of specialists. She set up a card catalog of Paleozoic foraminifers. In addition to an initial description of the species, as it is adopted worldwide, her catalog has also accumulated subsequent detailed descriptions that offer a complete view of any species. The data of this catalog can easily be processed by a computer for a quick and reliable determination of Paleozoic foraminifers. Rauser-Chernousova created an information center at GIN for the classification and stratigraphic study of Paleozoic foraminifers. She organized an exchange of work material between researchers by uniting them in a joint program. All these diverse forms of collective work naturally resulted in the creation of the Interdepartmental Micropaleontological Commission, which, until recently, Rauser-Chernousova was in charge of for more than 40 years. She used meetings, colloquia, and workshops to upgrade the study of foraminifers to solve different problems in geological practice. At present, many directors of geological institutions have lost the understanding that these meetings and colloquia represent the most favourable form for improving the experience of a researcher, and that they inevitably have a beneficial effect on the quality of geological work.

Rauser-Chernousova recognizes that the publication of articles is not only an exchange of information, but also an important tool for managing the process of collective research. Thus, she contributed much to putting together the first edition of *Voprosy Mikropaleontologii* in 1954. She has been a permanent editor of this publication for more than 40 years, and owing to the interest shown in the issue, she saw to it that the journal was translated into foreign languages. The worldwide recognition of this publication may be explained by the great consideration that Rauser-Chernousova has given it. Until recently, she read all papers submitted for publication, and all authors, regardless of their scientific degree and position, received her remarks, written in tiny and accurate handwriting and full of her benevolence and care. This does not mean that Rauser-Chernousova was a "good soul" as an editor. On the contrary, she was a rigorous and exacting editor; she often made an author rewrite an article in order to produce a distinct and clear text. Her demands of authors were very high, though individual in approach. She recruited many people for the sciences and helped them to choose the right path in their research.

The blossoming of micropaleontology occurred mainly because Rauser-Chernousova put what I call an integrated scientific-training-production complex into practice. Sometimes, this is called a "scientific school," but I do not believe that these terms are synonyms. The first term has a wider meaning, and it seems to be in use beyond the boundaries of the former USSR. I shall take a chance and suggest that, in the 21th century, these complexes will be among the most efficient ways of organizing the scientific process. To make such a complex work, it is essential to have not only the will, but also such features as humaneness and the desire to support and help in both life and practice.

It may seem that my thoughts are strange and bear no relation to the theme of this paper. However, they are unrelated only at first glance. The unified scientific-training-production complex is one of Rauser-Chernousova's most important ideas. This notion, along with her personal features, predetermined her success. My numerous meetings with Rauser-Chernousova have helped me to understand that she considers collective work to be the best form in which people can enjoy the wealth of their creativity; it is not simply a means for achieving rapid results. In other words, her wish was not to obtain, as has often been the case and continues to be today, but to give back, following the best traditions of Russian intellectuals.

I would like to emphasize that much of the success achieved in collective work depends not only on Rauser-Chernousova's organizational abilities and her concern for people, but also on the combination of these features and her pronounced creative personality and enthusiasm. A creative personality implies, first and foremost, constructive labour – a desire to do one's work best. Sometimes, we believe that coincidence or

simply chance is crucial to success. Even Rauser-Chernousova believes that, in her life, a decisive role in choosing the theme and location of her work was played by "... chance mainly. My contact with V.A. Dagaeva, a biologist at the Biological Station in Sevastopol', spurred me on to study oozes in Sevastopol' Bay. My meeting with V.N. Lyubimov accounted for my study of chlorophyll. On arriving in Moscow, V.A. Sulin proposed that I study the Paleozoic foraminifers of the Volga–Urals oil-bearing area. I was captivated by this work and obtained good results, and when the Geological Institute moved from Leningrad to Moscow, the director, A.D. Arkhangel'skii, invited me to join it. As the work conditions were good there, I agreed" (cited after Nalivkin, 1979, p. 162).

How simple appears the road that led Rauser-Chernousova to study foraminifers! She came from a wealthy family; her father was an engineer, and her mother, a music teacher. She was educated at the gymnasium and awarded the Silver Medal. Her father was opposed to her desire to enter the Higher Women's Courses, but her grandfather encouraged her. After she graduated from the Courses in 1918, she participated in an expedition to the Pechora region, which was interrupted by British intervention. Later, she worked in the Executive Committee of the Soviets. She married and gave birth to her daughter Irina. When her husband fell ill, they moved to the Crimea. Rauser-Chernousova resumed her scientific work in 1925 by studying the Carboniferous Ammonoidea from the South Fergana region and *Cardium edule* variability. As a young girl just graduated from the Courses, she possessed certain merits that prompted A.A. Chernov to offer her a job at the Pechora Expedition and Sulin to propose that she study Paleozoic foraminifers. Amongst these merits can be her persistence and determination to carry a work through to completion, which she exhibited in her first scientific work on Carboniferous Ammonoidea. These Ammonoidea were first described in her graduation thesis, but the description was published only 10 years later, in 1928.

For young researchers, the first step down the path of their career is the most difficult. The subsequent efficiency of their scientific activity will depend on this first step. For Rauser-Chernousova, the first step implied a choice of the method by which she would sample the material: either to collect the samples in a formation-by-formation method, as many researchers did and continue to do, or to choose another way. Dagmara Maksimilianovna began to collect fossils layer-by-layer, with an exact positioning of every sample. It is hard to tell now what prompted her to make this decision, whether it was intuition or a conscious realization of respect for "Mr. Fact," which was cultivated at the High Women's Courses. Regardless, she made the right choice, and why she did so will be a question for historians and philosophers of science to solve.

The layer-by-layer collection of foraminifers provided valuable material for research, i.e., a great number of facts, all with a fixed position according spatial and temporal coordinates, which maintained the sampling validity regardless of the changing views on the volume and age of a formation. This layer-by-layer collection of fauna helped Rauser-Chernousova to correct the "error of the century," i.e., to prove that the age of the top beds of the Carboniferous limestones in the Urals is similar to that of the basal layers of terrigenous sequences, then assigned to the Artinskian Stage of the Permian. The methods of revealing the succession of foraminiferal assemblages, based on layer-by-layer collection, helped her apprentices and disciples to solve disputed problems in stratigraphy. This success, as well as her elaboration of Carboniferous and Permian zonation, posed Rauser-Chernousova among the leading stratigraphers. Her works on zonal schemes are very popular. I would like to recall the method of superfine subdivision of sequences. This modest-sized work of Rauser-Chernousova is especially important now, when we should be paying special attention to increasing the quality of prospecting surveys, including their stratigraphic substantiation. The bulk of material was obtained through layer-by-layer sampling. How to synthesize this material is a matter of fundamental significance. The first step in synthesizing the material involves revealing the species categories. This seems quite natural, but the simple and clear principles elaborated by Rauser-Chernousova aided in defining the notion "species" for Paleozoic foraminifers.

The next step implies a systematization of taxa of superspecies rank for Paleozoic foraminifers, which requires a well-reasoned and logically irreproachable conception. Nalivkin characterized it in the following way: "... I often watched how Dagmara Maksimilianovna wrote her papers and envied her, because she knew how to present the material in a consequtive order and more completely, going from simple to complex, resisting any thematic temptations" (Nalivkin, 1979, p. 156).

In my opinion, the essence of Rauser-Chernousova's conception of classification is, in Vernadsky's terms, an empirical generalization. Empirical generalizations require special attention to terminology usage, and Rauser-Chernousova placed particular emphasis on this. Which is exemplified in *Terminologicheskii Spravochnik po Stroeniyu Rakovin Foraminifer*, which she compiled with Gerke (Rauser-Chernousova and Gerke, 1971). This reference book is only one part of a major work on terminology used when describing Paleozoic foraminifers. Unfortunately, this work has not yet been published in its full volume.

Micropaleontologists ignored empirical generalizations when discussing classification. About 10 years ago, I became aware of variations in the opinion of Rauser-Chernousova and her opponents. They believed that the hierarchy of taxonomic features should be

first established, by determining the features of species, genus, and family rank, etc. and then by logically elaborating the classification of superspecies taxa. This brings up the following question: Where did the hierarchy of taxonomic features come from? Different explanations are often presented to substantiate the hierarchy adopted *a priori* (biological implications, informational content of a taxonomic feature, and others). Such explanations represent only the fine cloths designed to conceal the *a priori* character of the selected hierarchy of taxonomic features. In my opinion, the second approach, which, I think, was of Rauser-Chernousova's style, comprises a sequential generalization of the empirical material on the variability of foraminifers. In fact, this approach also exploits axioms (the law of recapitulation and others) but these laws, on their own, are an empirical generalization of a huge scope of material. Applying these laws along with geochronologic and paleobiogeographic data, one can determine the position of microfossils in phylogenetic lineages. The latter helps to substantiate the hierarchy of taxonomic features. Rauser-Chernousova made the most important inference that the taxonomic range of a feature varies in the course of evolution. I really do not mean to ascribe my thoughts to her, as in some earlier works she gave credit to "biological significance" as well (for instance, by elucidating the wrinkled patterns of the shell wall when distinguishing the *Rugosofusulina* genus).

It was this approach to classification that led Rauser-Chernousova to the concept of periodicity and stages in the evolution of Paleozoic foraminifers. Having considered this problem in detail, she pointed out with good reason that periodicity reflects direct relationships between an organism and its environs, whereas stages represent more complicated bios-environs relationships. This discovery, made first in paleontology, is looked upon with new insight these days. Over the past 10 - 15 years, considerable attention has been given to the evolution of self-developing systems, which include humans. Their evolution is characterized by certain critical points. To survive, humankind should understand regularities in the passage of the self-developing system through these critical points. Ecosystems also belong to self-developing systems which passed repeatedly through such points, though we have yet to learn how this occurred in the past. It is precisely the study of periodicity and stages, based on the layer-by-layer collection of microfossils, that may offer some understanding. The strategy of survival of humankind in the 21st century, which was discussed at the UN Conference in Rio de Janeiro in 1992, is based exactly on elaborating rules of how to pass through the critical points. It is not inconceivable that, in the coming century, the study of periodicity and stages in the evolution of foraminifers will suggest an answer to this vital problem.

Rauser-Chernousova was born at the end of the 19th century. She is bound up with it by thousands of threads trough her education, culture, and scientific traditions. However, what she has achieved belongs to the 20th century. Much of her work will serve as a foundation for our activity in the 21st century. Her life and scientific activity are prominent examples of relationships not only between generations, but also between centuries.

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Micropaleontological Research of the Mesozoic and Cenozoic at the Geological Institute, Russian Academy of Sciences

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Abstract – The paper gives a review of micropaleontological investigations of the Mesozoic and Cenozoic that have been carried out at the Laboratory of Micropaleontology of the Geological Institute, Russian Academy of Sciences (GIN RAS) since the 1930s when the laboratory was established. It is shown how the spectrum of problems under study and the researched fossil groups and geography changed and expanded. The major stages in these investigations are defined.

Key words: **Mesozoic, Cenozoic, Jurassic, Paleogene, Neogene, foraminifers, nannoplankton, radiolarians, diatoms.**

Dagmara Maksimilianovna Rauser-Chernousova is the founder of the Laboratory of Micropaleontology at the Geological Institute, RAS. Initially, it was a "study of microfauna" within a division of stratigraphy; however, it soon was converted into a laboratory and became independent. Now, it is one of the few laboratories that, despite repeated reorganizations, has preserved not only its name, but also its unity and devotion to its scientific goals.

The scientific interests of Rauser-Chernousova and her colleagues primarily concentrated on Upper Paleozoic foraminifers, including their morphology, systematics, phylogeny, stages of development, and paleoecology, as well as an elaboration of zonal scales for the Carboniferous of the European part of the USSR.

However, Rauser-Chernousova realized that the development of Paleoroic foraminifers cannot be comprehensively studied without tracing their entire evolution. Thus, research cannot be restricted to Paleozoic groups alone, which are, moreover, quite peculiar and highly specialized. Investigations on Mesozoic and Cenozoic microfauna started at GIN in the 1950s on the initiative of Rauser-Chernousova, though Paleozoic foraminifers remained the main subject of the laboratory investigations for a long time. The first specialists on Mesozoic-Cenozoic microfauna appeared at GIN: V.G. Morozova (Upper Cretaceous-Paleogene), M.Ya. Serova (Paleogene-Neogene), T.P. Bondareva (Paleogene), V.A. Krasheninnikov (Paleogene-Neogene), and K.I. Kuznetsova (Jurassic). Initially elaborating all the scientific trends outlined by Rauser-Chernousova, they gradually turned their attention to working out a detailed zonal stratigraphy and correlation. This tendency has always been, and remains today, the prevailing objective in laboratory research.

The spectrum of stratigraphic research and their geography widened at that time. In the 1950s - 1960s, investigations were mainly focused on marine sediments of epicontinental basins of the USSR (the Russian platform), in the Crimea-Caucasus region, Central Asia, the Far East, and the eastern Mediterranean area. The main objective at that time was to elaborate and provide more details to foraminiferal scales.

A new stage in micropaleontological research of the Mesozoic and Cenozoic started in the late 1960s and beginning of the 1970s. It was characterized by the appearance of abundant and essentially new material on deep-sea drilling: multimeter continuous sections of slightly lithified sediments, almost completely composed of well-preserved microfossils. No sections of this kind are known from epicontinental basins. The deep-sea drilling project also involved biostratigraphers, who, in addition to traditional foraminifers, study other groups of microfossils, such as nannoplankton, diatoms, and radiolarians, as it became evident that detailed stratigraphical research is impossible without these fossils.

Since the 1970s, broader objectives have been advanced in micropaleontological studies on the Mesozoic and Cenozoic of epicontinental basins at the Geological Institute. The characteristic feature of this stage of stratigraphic investigations at GIN was an awareness by the majority of scientists of two important facts: (1) reliable stratigraphic interpretations cannot be obtained using only traditional macrofauna; data on microfauna should also be taken into account; and (2) for a comprehensive micropaleontological substantiation of the stratigraphic subdivision and correlation of sediments, other microfossils should be used along with foraminifers. In the 1970s, studies of nontraditional microfossil groups began in other laboratories of GIN.

By the early-middle 1980s, zonal scales using major microfauna and microflora groups had been essentially elaborated. Thus, they were no longer objects of study,

but became the subjects. Scientists at GIN could perform their research in areas with sections that are well-subdivided according to microfossils: the Jurassic and the Paleogene-Neogene of the Crimea–Caucasus region and the eastern Mediterranean region and the Paleogene of Central Asia. Though it is not very detailed, a reliable zonal stratigraphic scheme has been elaborated for the Paleogene-Neogene of the Far East and northeastern Russia, as well as for certain Neogene intervals of the Crimea–Caucasus region. The application of zonal scales it made possible to move on to a new, higher level of research: working out a scale of Mesozoic-Cenozoic events.

Now we shall dwell on some of the achievements made in the Mesozoic-Cenozoic micropaleontological investigations performed at the Geological Institute. In this paper, we can not afford to discuss in detail all of the vast information that has been obtained at present. Suffice it to say that specialists of the laboratory have published more than 30 monographs, including some that were published abroad, as well as many thousands of scientific articles and notes on the investigated subjects. We shall give a brief account of the results of investigations on major stratigraphic intervals, which have always been the focus of micropaleontological investigations at GIN (the Jurassic and Paleogene-Neogene), as well as the main regions (the Russian platform, the Crimea-Caucasus region, Central Asia, the eastern Mediterranean, the Far East and northeast of Russia, and the World Ocean).

The Jurassic. Research on Jurassic foraminifers started at GIN in the early 1950s. Classical Jurassic sections of the Russian platform, which contain rich foraminifer and ammonite assemblages, were the main subjects of investigations. Further, the research areas extended into Great Britain, Poland, France, and Germany. These materials served as a basis for a broad correlation of zonal subdivisions, identified by foraminifers. However, correlations with southern sub-Tethyan deposits still encountered difficulties. Therefore, in the 1970s, research on Jurassic foraminifers started in the Crimea, which resulted in an elaboration of the Jurassic zonal scheme based on this fossil group. In the 1980s, attention was concentrated on the eastern Mediterranean region (Syria), where the rich fauna of Tethyan foraminifers, previously unstudied, was collected and described. These investigations allowed a foraminiferal zonal scheme of the Jurassic to be made, and the main trends and regularities in the evolution of tropical foraminifers to be outlined; it also allowed for them to be correlated with the development of boreal fauna.

Along with foraminifers of the eastern Mediterranean, the laboratory deals with Jurassic ostracodes. They are a new group for Mesozoic studies in the laboratory and appear promising for both stratigraphy and paleoecology. The Jurassic zonal scheme of Syria was made on the basis of ostracodes and was correlated with the foraminifer-based scale for these sediments. A detailed study and description of the rich and diverse assemblages of Jurassic ostracodes of Syria greatly contributed to this work.

Research into Mesozoic sediments on the basis of deep-sea drilling materials greatly promoted investigations of Jurassic foraminifers. Legs 27 and 41 of D/V *Glomar Challenger* supplied data on the development of this fauna in the central Atlantic region and central part of the Indian Ocean. Foraminiferal assemblages composed of deep-water cosmopolitan species and devoid of any tropical Tethyan forms were discovered here. A similar dependance of the taxonomy of foraminiferal communities upon the bathymetry of a basin was also established for the Jurassic deposits of Syria. This allowed coeval Jurassic sediments in continents and oceans to be reliably correlated.

Sea-level changes in the Jurassic were investigated by analyzing the composition of foraminifers from deposits of transgressive and regressive phases of basin development both in the Boreal belt and in the Tethys. The problem of the synchronism and diachronism of these phases in seas of different latitudinal-climatic zones is at present, and will remain in the future, of particular interest, and it is a research subject that is especially emphasized by micropaleontologists at GIN.

Jurassic micropaleontological research carried out at GIN is characterized by one specific feature that is also related to Rauser-Chernousova's methods and style of work. It was Rauser-Chernousova who first applied the method of collective work, and GIN turned into an integrated center. The organized collective, headed by our Laboratory of Micropaleontological and the Vilnius Geological Institute, provided extensive materials on the Jurassic. We were all citizens of a great country, we were devoid of pragmatism and mercantilism. Nobody asked, "What benefits shall I get?", the question was rather, "What shall I contribute?". Therefore, vast material was gathered on the Timan–Pechora basin and the Crimea, Lithuania, Siberia, Ukraine, Moldavia, Byelorussia, Georgia, Central Asia, and the Moscow region. Based on these materials, we have elaborated a scheme for the zonal subdivision of the Jurassic by foraminifers, performed a regional correlation of the former USSR, and prepared several collective monographs and volumes of collected articles, which we continue to be proud of.

Turning to the work on Cretaceous and Paleogene deposits, we should primarily pay a tribute to V.G. Morozova, a prominent paleontologist who worked at our laboratory for more than 30 years. Her research on Maestrichtian-Danian foraminifers was of particular significance. Studies of microfauna and, on this basis, a zonal subdivision of the Danian Stage in the Crimea, which was the terminal stage of the Cretaceous at that time, made it possible to attribute it to the Cenozoic. Morozova possessed remarkable insight into morphology; following her work, nobody has described any new species or genera in the samples she

investigated or changed the volume of the taxa she described.

The Paleogene-Neogene. Micropaleontological investigations of this interval are intimately associated with the name of Valerii Arkad'evich Krasheninnikov, who succeeded Rauser-Chernousova as the head of the Laboratory of Micropaleontology. Krasheninnikov is a well-known authority in the field of the Cenozoic micropaleontology and stratigraphy. We shall merely list the problems and regions that he has investigated and is currently investigating: the Miocene of the southern part of the USSR; the Cenozoic of Siberia (in the context of these investigations, the method of mapping chronostratigraphic boundaries was applied for the first time in a geological survey); the Paleogene of the Crimea-Caucasus region; the zonal scale of the Paleogene (Krasheninnikov is one of its authors); the problem of the subdivision of the Cenozoic into stages; and, finally, the Cenozoic stratigraphy and oceanography of the World Ocean (Krasheninnikov was a participant on many legs of *Glomar Challenger*).

Due to the efforts of Krasheninnikov, Cenozoic micropaleontological investigations at GIN were gradually formed into a separate scientific line. In the 1970s, studies on carbonate nannoplakton, diatoms, radiolarians, and dinoflagellates began in the Laboratory of Micropaleontology and, later, in other laboratories of GIN. One of the strong points of the investigations on Cenozoic microfossils at GIN is their complex character. This can be seen in a series of works on the correlation of zonal scales for different Cenozoic fossil groups in continents and oceans.

The Crimea-Caucasus region, Central Asia, and their adjacent territories, where Paleogene deposits are widespread, are among the most interesting areas in the world. Here, the foundation of the foraminifer-based stratigraphy of the Paleogene was laid. The Paleogene of these regions has been always emphasized at GIN; therefore, they are annually visited by micropaleontologists. Recently, a foraminifer- and nannoplankton-based detailed subdivision of the Paleogene was done for Central Asia. The northern part (the Caspian Sea region and the areas of the middle Volga and middle Don rivers) is especially interesting, because, on the one hand, its Paleogene sequences are enriched in siliceous microorganisms, and, on the other, they contain interbeds of carbonate rocks with foraminifers and nannoplankton. This enabled micropaleontologists from GIN to make a more precise zonal scale for the Paleogene epicontinental basins of the eastern Peritethys according to diatoms and radiolarians (the basis of the diatom and radiolarian scales was elaborated by specialists of the Leningrad school) as well as to correlate them with the scales based on carbonate plankton.

Among recent investigations on the Neogene of the Crimea-Caucasus area and Central Asia, those pertaining to nannoplankton should be particularly emphasized. On the basis of these investigations, we have proved that over the most part of the Neogene, the basins of the eastern Paratethys had a constant, though loosened connection with the global ocean. Thus, the idea that the eastern Paratethys was predominantly isolated during the Neogene turned out to be incorrect.

Micropaleontological research on the Paleogene-Neogene and Quaternary sediments in the Pacific, Indian, Atlantic, and Southern oceans was based on various microfossil groups: foraminifers, nannoplankton, diatoms, radiolarians, and silicoflagellates. They comprised different paleotemperature belts, from the North Atlantic to the North Pacific regions in the north to the Antarctic areas in the southern hemisphere.

As a result, a great contribution has been made to studies on the Cenozoic stratigraphy of the World Ocean, and the major features of the sedimentary cover structure and the dynamics of its evolution in the Cenozoic have been clarified. Common zonal scales for oceanic and pelagic sediments on continents were elaborated; this provided opportunities for "reading" the geological history of continents and oceans with the same methodical techniques.

Micropaleontological investigations in the Far East are intimately associated with the name of another specialist of GIN, M.Ya. Serova. She began her career by investigating Cretaceous Paleogene foraminifers in the Carpathians, but later focused her attention on the Cenozoic of the Far East (Kamchatka, the Koryak Upland, Sakhalin, and other areas). The current concepts of the zonal stratigraphy and correlation in this region are, to a great extent, a result of the many years of research carried out by Serova. Beginning in the 1970s, siliceous microfossils (diatoms and radiolarians) and, later, calcareous nannoplankton were widely used by specialists from GIN for Cenozoic investigations in the Far East. The use of these microfossils allowed the Neogene sequence to be subdivided on a zonal basis and Paleogene stratigraphy to be made more precise.

Finally, the elaboration of a scale for Paleogene-Neogene events started in the 1980s. The work was done best for the Paleogene of the Crimea-Caucasus region and for Central Asia. In particular, it was proved that the sedimentation process in these regions was mostly controlled by eustatic fluctuations. It was established that the anoxic sedimentation intervals and ash/benthonite interlayers are related to eustatic transgressions and can serve as ideal datum levels.

In bringing this brief account of micropaleontological research on the Mesozoic and Cenozoic at the Geological Institute, RAS, to a close, we would like to mention the following. We all know that Russian science is experiencing hard times now. Many have adopted the mentality that we need to survive, and that we must work in order to survive. However, the motto that Dagmara Maksimilianovna Rauser-Chernousova has maintained throughout her long and productive career remains a challenge to us all: we do not work in order to survive, but we live in order to work and create.

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Facial-Ecological Structure of the Derevnya Formation Microbiota: Upper Riphean, Turukhansk Uplift, Siberia

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Abstract – The facial-ecological and microphytological features of the siliciclastic-carbonate Upper Riphean Derevnya Formation in the Turukhansk Uplift imply that it accumulated in sea shelf environments. The dominating carbonate complexes were formed in the outer open shelf settings, and the siliciclastic deposits were formed in the inner protected shelf. The Derevnya deposits yielded 44 genera and 53 species of organic-walled microfossils, found at 11 stratigraphic levels. Among them, *Chuaria*, acanthomorphic acritarchs, specific coccoid forms, large ellipsoid envelopes, broad smooth-walled and wrinkled sheaths, septate filaments and thalluses (including branching ones), are of particular importance. The composition and facial-ecological setting of each identified microfossil assemblage are analyzed. Bathymetric control was the primary factor affecting the microbiota lateral structure, which was also under the influence of hydrodynamic factors and secular trends in the environmental evolution.

Key words: **Organic-walled microfossils, microbiota, the Riphean, the Proterozoic, facies association, paleoecology, shelf, paleoenvironments, carbonate platform.**

A comprehensive understanding of the Precambrian biosphere requires detailed studies of both the organisms of that time and their environments. These two lines of biospheric investigations have long been evolved either separately or in combination. Thus far, only scarce examples of their synthesis have concerned carbonate deposits with enclosed phytoliths and/or silicified microfossils (e.g., Walter, 1976; Schopf, 1983; Schopf and Klein, 1992). Attention was focused on various aspects of the specific carbonate sedimentation and local paleoenvironments during the early silification of microorganisms. A more representative and diverse component of the Precambrian biota, such as organic-walled microfossils, appeared to be out of consideration. However, microorganisms of this preservation type are associated with a wider range of paleoenvironments than silicified microfossils; therefore, they are more indicative of the early stages of the biospheric evolution. There are still few facial-ecological reconstructions of the Precambrian paleoenvironments based on organic-walled microfossils (Vidal and Nystuen, 1992; Zang and Walter, 1992; Hofmann and Jackson, 1994; Veis and Petrov, 1994a, 1994b), although the great potential for obtaining new paleoecological information in the near future is quite evident. As a result of our previous works (Veis and Petrov, 1994a, 1994b), a general distribution pattern of microfossil assemblages in the Riphean epicratonic basins of Siberia was revealed, showing the strict dependence of the taxonomic diversity and abundance of microbiotas on the accumulation depth of the enclosing deposits, and several compositionally specific microfossil groupings were identified: deep-water, medium-water (optimal), transitional, and shallow-water ones. This phenomenon was interpreted to reflect the original paleoecological distribution of microorganism assemblages.

The objectives of this study are as follows: (1) to analyze the facial-ecological distribution of taxonomically different organic-walled microfossil assemblages in the shelf siliciclastic-carbonate Upper Riphean sediments; (2) to refine and substantiate the distinguished (Veis and Petrov, 1994a, 1994b) distributional patterns of taxonomically different microbiotas, using new types of sediments as examples; and (3) to define the possibility of solving these problems more adequately, first of all, with regard to the microorganisms from the "optimal" zone of the paleobasins. The siliciclastic-carbonate deposits of the Derevnya Formation of the Turukhansk region, northwestern Siberia, which enclosed abundant and diverse organic-walled microfossils, were the object of the study.

FACIES AND DEPOSITIONAL ENVIRONMENTS OF THE DEREVNYA FORMATION

Age and composition of the Derevnya Formation. Among the predominantly carbonate Riphean deposits of the Turukhansk region, the Derevnya Formation is distinct due to several features, such as the prevalence of variegated stromatolitic dolomites and the lesser abundance of limestones, the small-scale cyclicity in the vertical succession of stromatolite morphotypes, and the occurrence of two siliciclastic members that are also variegated. The formation is traceable over a distance of 100 - 110 km from the north to the south within two blocks (monoclines) bounded by regional thrust-faults (Fig. 1). The most complete

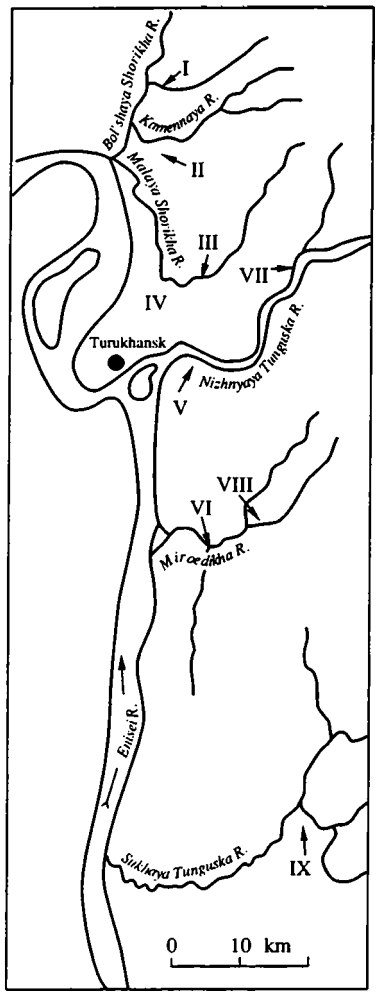

Fig. 1. Geographic location of the Derevnya Formation sections.

I – Nadporozhnaya R.; II – Kamennaya R.; III – Malaya Shorikha R.; IV and V – Nizhnyaya Tunguska R.; VI – Miroedikha R., the western monocline; VII – Nizhnyaya Tunguska R.; VIII – Miroedikha Vtoraya R., the eastern monocline; IX – Sukhaya Tunguska R.

sections of the formation are known in the western monocline along the Nizhnyaya and Kamennaya Tunguska rivers. Other sections exhibit only some parts of the formation, composed mainly of carbonate deposits; siliciclastic members are either unexposed or inferrable from scattered exposures and elluvial debris of claystones, siltstones, and sandstones.

The Derevnya Formation is assigned to the Upper Riphean on the basis of the stromatolite and microfossil taxonomic composition (for review and bibliography, see Semikhatov, 1991). Among microfossils, acanthomorphic acritarchs *Trachyhystrichsphaera* and *Prolatoforma* (unknown in the region prior to our investigations) are worth particular attention (see below).

The available K-Ar datings of glauconite (820 - 870 Ma; Ivanovskaya *et al.*, 1973) show the minimal age limit of the formation only. Its maximal limit (1017 ± 90 Ma) is determined by the Pb–Pb isochron dating of carbonates of the underlying Sukhaya Tunguska Formation (Ovchinnikova *et al.*, 1994). The Derevnya sediments overlie the clastic and pelitomorphic dolomites of the latter formation with a regional unconformity as follows from the boundary relief, the erosional forms of which have an amplitude of up to 40 cm. Upsection, the Derevnya sediments grade into fine-grained limestones and dolomites of the Burovaya Formation.

The Derevnya Formation can be arbitrarily subdivided into three subformations (Fig. 2). The lower subformation (ranging in thickness from 105 to 115 m) is composed of pale-yellow or red stromatolitic and sparse clastic dolomites, which include some thin (up to several tens of centimeters) shale interbeds. The middle subformation (100 m thick) comprises two siliciclastic members separated by a unit of stromatolitic dolomites and subordinate limestones. The siliciclastic members are dominated by shales and siltstones, which are separated from the underlying stromatolitic dolomites by a sharp boundary and are overlain conformably by carbonate members of the succession. The upper subformation (up to 70 m thick) is composed of grey clastic, pelitomorphic, and stromatolitic dolomites. In an east–west direction, the thickness of the Derevnya Formation increases noticeably from 75 to 270 m, whereas in a south–north direction, it increases from 220 to 270 m. The lower and middle subformations are the most variable in thickness (Fig. 2).

The organic-walled microfossils are confined to the lower and middle subformations, both including carbonate-free shales. In this paper, we will focus our attention mainly on the siliciclastic facies of the Derevnya Formation, which yield microfossils.

Facies associations of carbonate successions. The Derevnya deposits demonstrate three carbonate facies associations (Fig. 2). The most widespread of them, especially in the lower and middle subformations, is the association of stromatolitic biostromes, which reveal a cyclic alternation of different buildups: abundant *Conophyton* and *Jacutophyton*, less frequent *Baicalia*, and rare *Inzeria* and *Colonella*. The alternation of the first three stromatolitic morphotypes forms the so-called "*Jacutophyton* cycles" (Serebryakov and Semikhatov, 1974; Serebryakov, 1975), which range in thickness from dozens of centimeters to several meters. The composition and structures of the intercolumnar carbonates within these biostromes suggest that the lower parts of the biostromes, consisting of *Conophyton*, represent quiet-water paleoenvironments below the storm wave base, while their upper parts, composed of *Jacutophyton* that grade upward into *Baicalia*, represent the sediments of the permanent wave activity zone. Thus, this facies association was formed in an open-sea setting, including its relatively deep zone, characterized by frequent sea-level oscillations.

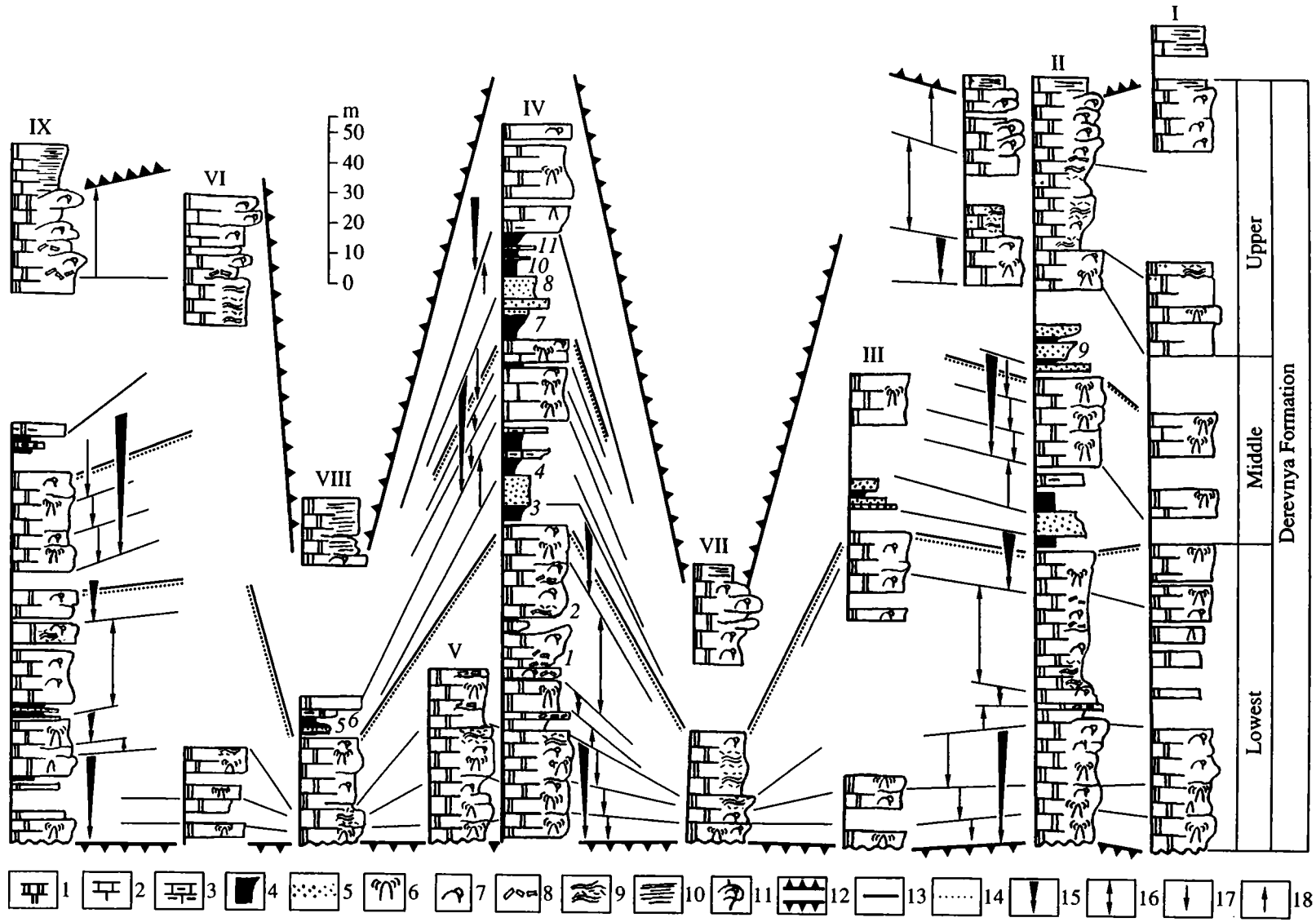

Fig. 2. Structure and facies of the Derevnya Formation.

1 - 5 – lithological composition: 1– dolomites, 2 – limestones, 3 – clayey limestones and marls, 4 – shales and siltstones, 5 – sandstones; 6 - 9 – facies associations of carbonate successions: 6 – the association of rhythmically composed biostromes, 7 – the association of bioherms and biostromes, 8 – the coarse clastic type of the dolomite association, 9 – the fine clastic type of the dolomite association; 10 – horizontally bedded carbonates of the Burovaya Formation; 11 – large stromatolite bioherms; 12 – the top and base of the Derevnya Formation; 13 – boundaries between the subformations; 14 – bases of the siliciclastic members; 15 – regressive successions; 16 – compound successions; 17 – regressive trends; 18 – transgressive trends. Hereinafter the numbers denote the section localities and levels with microfossils.

A less significant constituent of the formation is an association of bioherms and biostromes of columnar branching stromatolites, which do not show the cyclic morphotype alternation. This association is widespread in the upper and lower subformations, where it forms packages that are up to 10 m thick, composed of *Baicalia* and rare *Inzeria* and *Gymnosolen*. Horizontally- and cross-laminated clastic carbonates dominating in the intercolumnar and interbiohermal spaces and the frequent lateral replacement of various mechanogenic sediments are indicative of variable hydrodynamically active settings. In general, this association was formed in the permanent wave activity zone.

A clastic dolomite association containing sporadic stromatolites is a minor component of the formation. This association unites deposits of several facial types, two of which are the most common. The first type is represented by an alternation of horizontally- and wavy-laminated dolarenites, fine flat-pebbled breccias, and pelitomorphic dolomites. The second type is distinguished by wackestone beds, in which the flat unrounded flakes, up to 10 cm long, tend to have oriented along the cross-laminated intervals with amplitudes of up to a few dozen centimeters. The deposits of the first type are confined to the top of large-scale shallowing cycles. Their structures and positions in the sections suggest low-energy, very shallow water environments, typical of zones from tidal flats to shallow lagoons behind reefs. The deposits of the second type are spatially associated with bioherms of columnar branching stromatolites and thin shale interbeds. They occur in the basal parts of the transgressive trends (Fig. 2) and were formed under high wave activity conditions during the rapid flooding of the flattened shallow-water landscapes.

Facies associations of siliciclastic members. Five facies can be recognized in the sandy-clayey carbonate-free deposits of the Derevnya Formation (Fig. 3).

(1) The shale facies comprises relatively homogenous clayey deposits of greenish-grey, black, and cherry-red colors with even and fine horizontal bedding. The bedding is emphasized by thin (less than 0.1 mm) films of organic matter, which probably represent traces of benthic microbial assemblages (mats), common components of many Proterozoic shaly units (Fairchild and Herrington, 1989; Schieber, 1990). The red color of the deposits is a result of the occurrence of scattered hematite, which is not always distributed along the bedding planes, and seems to be related to later diagenetic processes. Some laminae of the shales are marked by an admixture of silty quartz grains, but traces of erosion are not present at the bases of these laminae. This suggests that they appeared as a result of the slow precipitation of siliciclastic material from suspended flows. The deposits of the shale facies constitute homogenous, relatively thick, laterally continuous packages in the upper parts of both siliciclastic members of the formation. Their lower boundaries are sharp, with a smooth erosional relief in some places. Upsection, the packages grade into clayey-carbonate deposits. The above implies that the shale facies sediments accumulated in quiet and relatively deep-water settings.

(2) The siltstone facies is composed of horizontally bedded clayey silts, which contain sporadic thin beds (0.5 - 3.0 cm thick) of horizontally and cross-bedded fine-grained sandstones. The bedding of the clayey-silty packages is emphasized by the distribution of fine-grained quartz sand and, in some places, by glauconite grains associated with small-scale local erosional surfaces. More apparent erosion traces are observed at the base of sandy interbeds, which comprise numerous inner reactivation surfaces that are indicative of local erosion and, in some places, overlain by thin (up to 2 mm) shaly laminae. The cross-bedding sandstone, which shows no oscillatory structures, suggests the deposition of sandy particles under unidirectional bottom-current conditions. The deposits of this facies constitute small packages, which grade upward into sediments of the shale facies or are overlain by the coarse-grained sandstone beds. These features imply that the deposits of the siltstone facies accumulated in more proximal and shallower settings than those of the shale facies. The silty and sandy matter was deposited from suspended overbottom and loaded bottom unidirectional flows outside the wave zone. The sediments seem to have been transported mainly by tidal currents within the lower subtidal zone.

(3) The sandstone facies is represented by beds (0.5 - 2.0 m thick) of light-grey, black and brown, medium- to coarse-grained quartz arenites containing rare lenslike sandy-gravel accumulations. The grain size varies chaotically within a bed, but the coarsest grained fractions tend to be confined to the upper parts of the beds. The sandstones are characterized by massive or, less frequently, uni- and multidirectional cross-bedding of variable amplitudes (2 - 20 cm), with diverse dip angles and frontal surface forms. Some sandstone horizons show cross- and cross-wavy-bedded ripple marks with an amplitude of 10 cm. The most characteristic feature of this facies is an abundance of inner reactivation surfaces overlain by discontinuous shaly laminae (0.5 cm thick). This results in the characteristic flaser lamination of the sandstones, with sometimes small (1 - 5 cm) lenticular jointings.

The specific bedding and structure of the sandy units indicate that they were formed by unidirectional loaded flows. It is significant that the structure succession exhibits no signs of long-period oscillatory or combined-flow activity. The coarsest grained sandy-gravel units confined to the erosional surfaces are probably reworked and redeposited sediments. Some horizontally bedded structures of the sandstone facies reveal a similarity to transflow bedding, which was described in detail earlier (Allen, 1982). Most of the flasers were evidently formed under a reversal-stage regime. Thus, the sediments of this facies originated

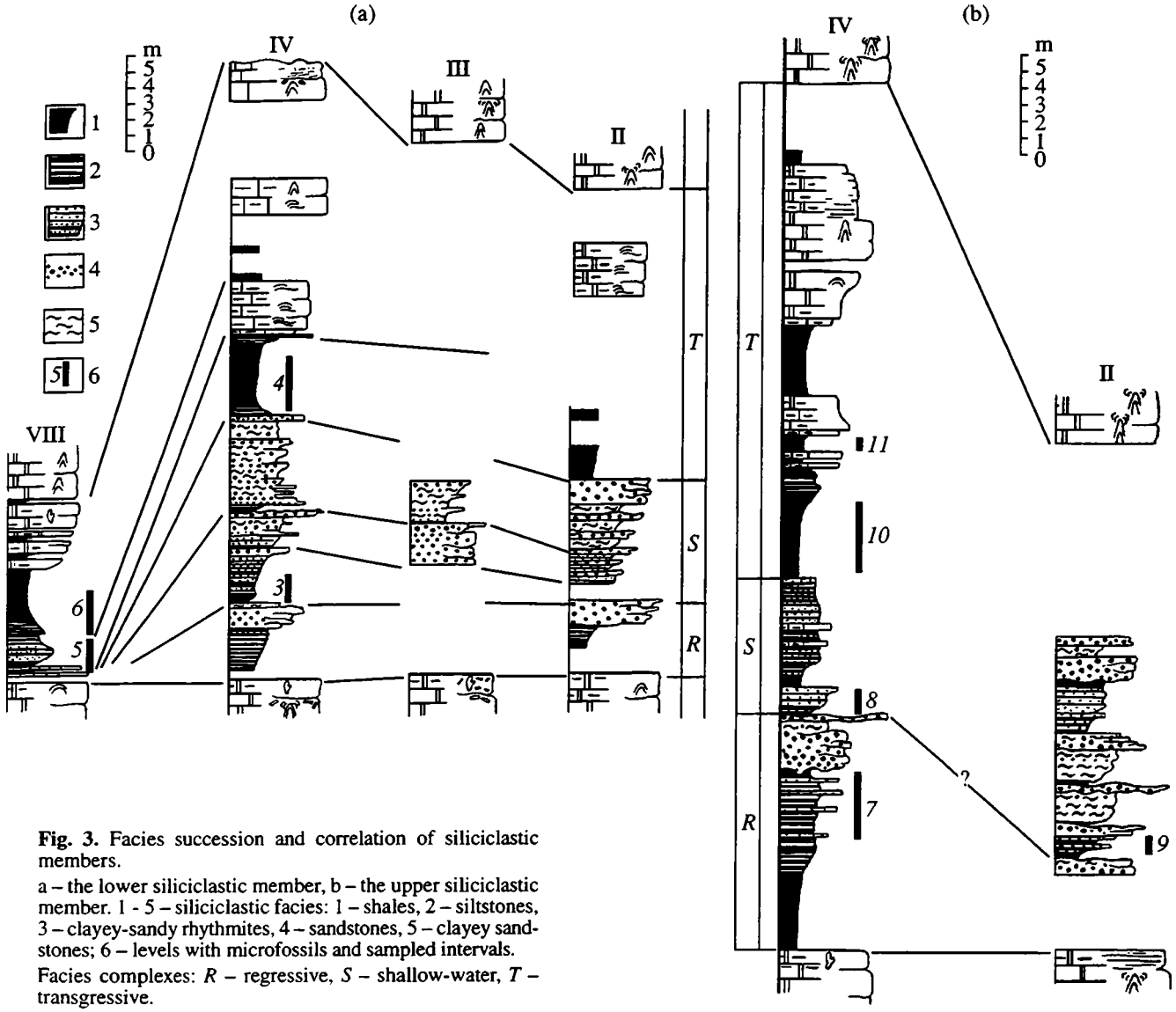

Fig. 3. Facies succession and correlation of siliciclastic members.

a – the lower siliciclastic member, b – the upper siliciclastic member. 1 - 5 – siliciclastic facies: 1 – shales, 2 – siltstones, 3 – clayey-sandy rhythmites, 4 – sandstones, 5 – clayey sandstones; 6 – levels with microfossils and sampled intervals.

Facies complexes: *R* – regressive, *S* – shallow-water, *T* – transgressive.

under conditions of periodic unidirectional flows alternating with episodic quiet-water conditions marked by clayey matter accumulation. Such conditions might occur in an upper subtidal–lower intertidal zone, i.e., in a peritidal setting favorable for the accumulation of sandy units (Klein, 1977; Clifton, 1983).

(4) The clayey sandstone facies is represented by drastically unsorted sandy-shale deposits that are brown and greenish-brown in color. They constitute packages, up to 2 m in thickness, in which sediments with a dominating clayey component (quartz wackes) grade gently or chaotically into poorly sorted sandstones with abundant clayey-silty matrix. The granulometric composition of the rocks tends to be of the bimodal pelite-psammite distribution. Embedded pyrite crystals are confined to more clayey varieties. The deposits are horizontally bedded; some sand beds and lenses have erosional lower boundaries.

The most clayey deposits of the facies under consideration demonstrate peculiar fragmental postsedimentary microstructures, evidencing disturbed sediment tenacity and intralayer displacements. They are similar to pseudonodular structures (Klein, 1971, 1977), which resulted from the combined dilution and loading of sediments under conditions of extremely early sediment consolidation, or to structures related to repeated periods of subaerial exposure and indicative of the initial soil-forming processes (Wright *et al.*, 1986).

This facies is closely associated with the sandstone facies, being overlain or laterally replaced by the latter. The facies was formed in shallow-water environments with some probable periods of subaerial exposure. The current velocities were insufficiently high to form the condensed deposits or debris flows forming the cross-bedded sediments. The sediments of this facies probably accumulated in semiisolated bays or lagoons under the influence of periodic tide currents. Their probable

analogs are known among the recent microtidal deposits of West Africa (Fleming, 1988).

(5) The facies of sandy-clayey rhythmites is represented by frequent alternations (within 15 - 30 cm intervals) of grey, predominantly clayey packages with interbeds of fine- and small-grained sandstone. The packages also include some persistent laminae enriched in variable amounts of sand. The deposits usually demonstrate even horizontal lamination, and only individual sand beds have a flat parallel cross-bedded series. In the most clayey deposits (the upper siliciclastic member at the Nizhnyaya Tunguska River), there are 0.5 m thick isolated beds of dark-grey limy shales and clayey limestones; their horizontal bedding is emphasized by films enriched in organic matter. The amount and grain size of the sandy matter increase noticeably upward in the section, and this facies is gradually replaced by the sandstone facies.

The sandy-clayey rhythmites accumulated in quiet-water environments were affected by frequent injections of sandy matter. The facies position in the general succession of the Derevnya deposits suggests that it was formed in the upper subtidal zone, near the area of the sandstone facies distribution. The sedimentary material was transported exclusively by suspended flows over small distances. In the offshore direction, the sandy-clayey sediments grade rapidly into clayey-silty and, then, into clayey-carbonate deposits (the upper siliciclastic member is an example). The successive changes in the facies composition indicate the progradational regime of its formation.

Facies successions and paleoenvironmental changes in the Derevnya shelf. The above suggests that the Derevnya sediments were deposited on the siliciclastic-carbonate shelf. The relatively high submergence rate of the Derevnya time was typical of the northwestern part of the Turukhansk region, whereas its eastern part was positioned on the slope of the extensive consedimentary uplift, recorded first by Dragunov (1959). The thickness and facies distribution indicate that the open sea basin was located, at that time, to the northwest of the recent Turukhansk Uplift, being replaced northeastward by coastal landscapes. The carbonate and siliciclastic depositional environments in the basin were very different.

The signs of the active wave sedimentation regime that are widespread in the carbonate deposits suggest that they were formed mainly in the open-sea outer shelf. These deposits, dominating over the Derevnya Formation, compose either distinct regressive successions reflecting clearly pronounced shallowing trends or less distinct compound successions devoid of such trends.

The regressive successions show thick (up to several tens of meters) shallowing trends and are separated by sharp, frequently erosional boundaries. These successions are most widespread in the lower and middle subformations. Their lower horizons are constituted by *Conophyton-Jacutophyton* biostromes cyclic in structure; the middle horizons are composed of associated bioherms and biostromes, without a cyclic structure. The upper ones are composed of an association of clastic dolomites of the first type. The successions are laterally the most continuous members of the Derevnya Formation, and generally correspond to high-sea-level facies systems. They seem to reflect the following environmental changes. The association of cyclic *Conophyton-Jacutophyton* biostromes was formed during maximal flooding, and their characteristic inner cyclic variations in stromatolite morphology reflect the short-term sea-level oscillations. When the basin depth decreased, this laterally continuous association broke down into a mosaic complex of various stromatolite buildups, which were formed under intense and permanent wave activity conditions (the association of bioherms and biostromes without cyclic structure). A further fall in sea level resulted in the degeneration of the stromatolite buildups into small bodies of layered, less frequently columnar-layered morphotypes alternating with fine-grained deposits (the first type of the clastic dolomite association).

Compound successions, devoid of a single evolutionary trend, are typical of the lower and upper subformations, where they are represented by relatively shallow-water facies associations. The successions show transgressive trends, up to 10 m thick, and subordinate small-scale shallowing trends. The first trends are characterized by a change upward from the clastic dolomites to the stromatolite varieties composed of columnar buildups, and the second ones, by the reverse facies change. None of the trends include cycles of the second order. The section intervals represented by these successions correspond to facies systems of low sea-level stand (Fig. 2). The carbonate deposits of the transgressive trends show thin (up to 5 - 10 cm) shale interbeds, some of which contain the organic-walled microfossils. These successions reflect the following events.

Predominantly coarse wackestones (the clastic dolomite association of the second type), which are confined to basal horizons of the successions, and the isolated stromatolite bioherms (*Baicalia*) correspond to the initial stages of transgressions, the landscapes of which, resulted from the rapid flooding of a flat peritidal zone. A noticeable increase in the abundance of stromatolites, first of biostromes and later of large bioherms (*Inzeria*, *Baicalia* and sometimes *Jacutophyton*), characterizes the advance of the transgression. The synoptic relief of these bioherms reached several meters and protected the quiet-water interbiohermal space, which sometimes served as a trap for transitional flows of fine clayey material from distant basin areas. Clayey siltstones of one large trap contain a microfossil assemblage that is unique in its preservation and diversity (see below). Thus, the described sedimentary successions reflect paleolandscape changes from shoals and protected lagoons to flat shallow settings occupied by stromatolites (mainly *Baicalia*) and, further, to relatively deep

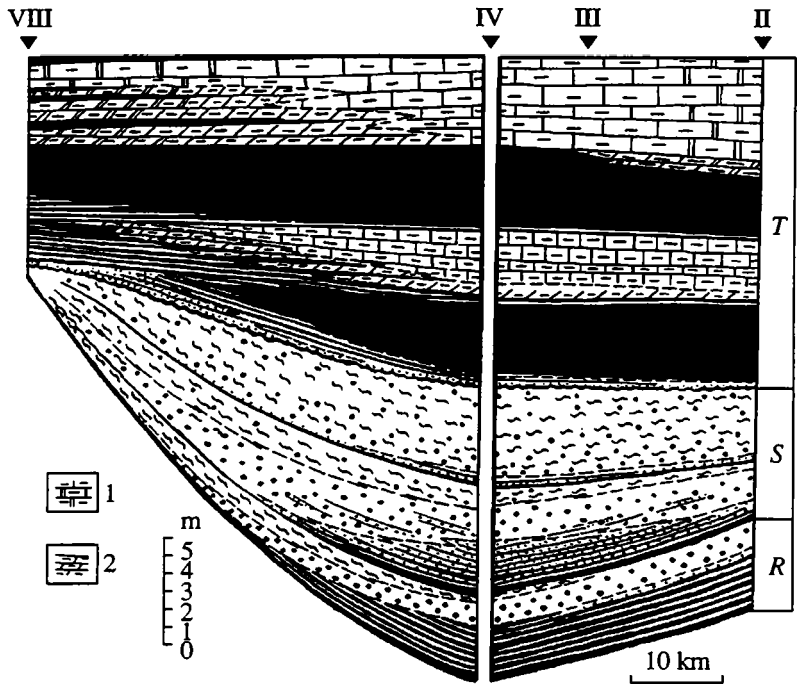

Fig. 4. Inferred facial structure of the lower siliciclastic member.
1 – clayey limestones or dolomites, 2 – marls (for other symbols see Fig. 3).

areas with the distinct relief of biohermal structures, which could serve as barriers.

Unlike the Derevnya carbonates, the two siliciclastic members of the middle subformation were formed in the inner (protected) shelf, as evidenced by the absence of any traces of a stormy sedimentation regime. The sedimentary material was transported here mainly by tidal currents. All changes in facies complexes and depositional environments were controlled by the sea-level stand and the amount of siliciclastic supply from the east. Each siliciclastic member shows three facies complexes (Figs. 3 and 4).

The lower regressive complex composed of siltstone and sandstone facies corresponds to a clearly pronounced shallowing-up trend. At the time of its deposition, considerable siliciclastic masses spread rapidly in the offshore direction, covering the entire shelf area as a result of the rapid migration of the siliciclastic front. Its approach is recorded by the rapid but gradual increase in scattered clayey admixture in dolomites underlying the siliciclastic members (Fig. 2).

The middle shallow-water complex, which is represented by a complicated alternation of sandy-clayey rhythmites, sandstones, and clayey sandstones, was formed in the most proximal and shallowest water settings. These deposits rest on the underlying regressive complex and are separated from it by a sharp boundary. However, there are no signs of erosion at the contact, except for an intraformation unconformity recorded at the Miroedikha Vtoraya locality. The data presented in Fig. 4 suggest that the deposition of the middle shallow-water complex of the lower siliciclastic member occurred under a fast progradational regime, with the final accumulation of clayey sandstone and sandstone facies occurring in extremely shallow-water landscapes. These landscapes were comprised predominantly of very shallow bays and lagoons, which experienced microtidal influence and sometimes isolation because of the periodic emergence of mobile sandy bars. In contrast, the facies relationship at the Nizhnyaya Tunguska locality downstream from the Strel'nye Gory (Fig. 3) evidences a more distal depositional setting of the shallow-water complex of the upper siliciclastic member. Scattered beds of silty shales and clayey limestones alternating with the finest varieties of sandy-clayey rhythmites suggest that the pure siliciclastic sediments of this complex graded seaward into the clayey-carbonate deposits.

The upper transgressive complex is distinguished by a prevalence of shale facies containing the richest organic-walled microfossil assemblages of the Derevnya Formation. This complex developed against the background of three successive stages of shelf flooding (Fig. 4). Each stage was characterized by a retrograde regime of sedimentation. Thus, the clayey sediments of the upper complex reflect the progressing increase in the depth of the paleobasin. These deposits are preceeded by erosional surfaces and small beds of basal breccias (the lower siliciclastic member), or by a sharp boundary without any traces of erosion (the upper siliciclastic member of the Nizhnyaya Tunguska locality). These boundaries are probably the main transgres-

sive surfaces within the formation, ranging from the erosional one of the proximal shelf zone to the conformable boundaries of the basinal settings. The inner boundaries within the sedimentary successions, which separate the above stages of the flooding, and the top boundary of the entire transgressive complex represent the conformable transgressive surfaces. The transition from the upper transgressive complex of the siliciclastic members to overlying, predominantly stromatolite dolomites reflects the further transgression advance and the transition from inner shalf environments of clayey sedimentation to open shelf areas of carbonate sedimentation.

ORGANIC-WALLED MICROFOSSILS OF THE DEREVNYA FORMATION

General characteristics of the Derevnya microbiota. The organic-walled microfossils of the Derevnya Formation were first reported by Timofeev (1966, 1969), who described a small assemblage of simple sphaeromorfites from the Nizhnyaya Tunguska section. Later, Mikhailova (1985, 1986, 1987) and Veis (1986, 1988) provided new and expanded information on the microbiota composition and described a number of new localities. As a result, this microbiota appeared to be one of the most abundant and diverse Precambrian microbiotas of northern Eurasia. However, neither a detailed description of specimens nor the correct geographic and stratigraphic position of most indicative forms have been reported. Our studies supplement previously published data on the Derevnya microbiota composition and earlier inferrences about its lateral variations, and reveal the ecological control over the distribution of particular assemblages of the Derevnya microorganisms.

We analyzed 85 samples from three sections different in microfossil abundance: the Nizhnyaya Tunguska locality, downstream from the Strel'nye Gory (68 samples); the Miroedikha Vtoraya locality (14 samples); and the Kamennaya River outcrops (3 samples). Determinable microfossils of good and exceptionally good preservation were found in 52 samples (41, 9, and 2 samples, respectively). To obtain representative data, we collected samples from rock varieties that were most promising for microphytological analysis and, sometimes, from less perspective ones.

We did not restrict the sample maceration to the usual techniques of treating with concentrated hydrofluoric acid and centrifuging the insoluble residue in heavy liquid (specific weight 2.1 - 2.2 g/cm^3). Before centrifuging, the residues of most samples were calibrated with a sieve, the cell diameter of which, was less than 50 μm. The rest of the material was studied under the microscope to select microfossils. This approach, which has already been used by German (1974), Vidal (1976), Burzin (1989), and Veis and Vorob'eva (1993), allowed us to obtain some large intact complex morphotypes of fossils that were earlier unknown from the Derevnya biota. All the Derevnya microfossils are represented by organic-walled forms. No silicified microfossils were found (Golovenok and Belova, 1993).

The microfossils occur at 11 stratigraphic levels within the formation (Fig. 2). Each level is an interval of a section, from several centimeters to few meters in thickness, with a distinct microfossil composition. The levels are separated by rock packages or members barren of organic remains. The observed variations in microfossil composition within some levels reflect short-term environmental changes during the formation of the microfossil assemblage.

The richest and most diverse microfossil assemblages were found in the Nizhnyaya Tunguska section downstream from the Strel'nye Gory at eight stratigraphic levels (levels 1 - 4, 7, 8, 10, and 11 of the general succession). In two other sections, the microfossils occur at two levels (5 and 6) of the Miroedikha Vtoraya section and at one level of the Kamennaya section; the last level (a 6-m-thick member) was assigned to level 9 of the general succession according to lithostratigraphic correlation.

When classifying the studied forms, we identified, as earlier (Veis, 1988, 1993; Yankauskas, 1989), first the supergeneric morphologic categories, such as acritarchs, collonial coccoidal forms, filaments (trichomes and sheaths), and complex microfossils, and then the most common and customary taxa of the generic and species levels (Veis and Vorob'eva, 1993). This approach permitted us not to include many simple and unimpressive microfossils into the recent form system in several mutually exclusive positions (e.g., Sergeev, 1993). In addition, this minimized the destabilizing effect of reshuffling well-known taxa (Knoll *et al.*, 1991; Butterfield *et al.*, 1994). As a result, 44 genera and 53 species were identified in the Derevnya microbiota.

The most diverse and abundant forms are acritarchs (11 genera, 19 species; 45% of studied specimens) and sheaths (14 genera, 16 species; 25% of found forms). Trichomes and trichomelike structures are represented by seven genera and six species (7% of specimens), and collonial coccoidal microfossils, by eight genera and eight species (20% of specimens). The scarce elements of the biota include ribbonlike thalluses (including branching forms), fossil remains that are similar in morphology to the reproductive structures of some eumycete fungus, cuticular covers of supposedly wormlike organisms, and other complex structures (four genera, four species; 3% of specimens). Some microfossils represent new taxa.

In general, the Derevnya microbiota is a typical (although very diverse) community of the Late Riphean pro- and eukariotes with a characteristic predominance of phototrophic organisms. Benthic and planktonic microbes can be identified with different degrees of confidence. Below, we consider the microphytological

Plate I. Microfossils of the Upper Riphean Derevnya Formation of the Turukhansk Uplift.

All described specimens are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

Levels of the Derevnya microfossil occurrence:

3, 8 - 11, 15, 18 - 20, 22 - 24, 27 - 30 – The second level, the lower subformation (Nizhnyaya Tunguska River downstream of the Strel'nye Gory); *7* – the third level; *16* – the fourth level; two last levels in the middle subformation, the lower siliciclastic member; the same locality.

1, 6 – The fifth level; *2, 14, 26* – the sixth level; both in the middle subformation, the lower siliciclastic member (Miroedikha Vtoraya River); *25* – the ninth level; the middle subformation; the upper siliciclastic member (Kamennaya River); *5, 21* – the tenth level; *4, 12, 13, 17* – the eleventh level; both in the middle subformation; the upper siliciclastic member (Nizhnyaya Tunguska River downstream of the Strel'nye Gory).

Bar for *1, 3, 5, 16, 21, 29* – 250 μm; for *2* – 400 μm; for *4, 7, 15, 24, 30* – 150 μm; for *6, 12* – 300 μm; for *8, 11, 14, 18, 20, 27* – 120 μm; for *9, 10, 19, 28* – 100 μm; for *13* – 450 μm; for *16, 17* – 700 μm; for *22* – 500 μm; for *23, 25, 26* – 80 μm.

1 – *Trachyhystrichosphaera aimica* Herm.; sp. no. 4010-40.

2 – *T. sp.*; sp. no. 4014-91.

3, 29 – *Simia nerjuenica* A. Weiss, large marginated envelopes with distinct peripheral zone; *3* – sp. no. 4233-59, *29* – sp. no. 4231-65.

4, 12, 13, 17 – microfossils of specific structure consisting of two subspherical envelopes enclosing one another; the inner envelope is locally covered with numerous closely spaced thin processes; *4, 12, 13* – magnified fragments of the inner envelope (the areas with processes); sp. no. 1898-64.

6 – Ellipsoidal envelope with thin folds yielding a typical rhombic pattern when specimens are overlapped; sp. no. 4009-81.

7 – Accumulation of different-sized envelopes *Leiosphaeridia* spp.; sp. no. 2378-176.

8 – *Nucellosphaeridium minutum* Tim., a sublinear aggregate of envelopes with inner spots; sp. no. 1810-75.

9, 10 – *Simia annulare* (Tim.) emend. Mikh., small marginated envelopes with peripheral zones; *9* – sp. no. 1810-150, *10* – sp. no. 1810-151.

11, 18 - 20, 27 – *Stiganema*-like filamentous forms; *11* – sp. no. 1810-191; *18* – a fragment of the filament with true lateral branching, sp. no. 4233-61; *19* – sp. no. 1810-153; *20* – sp. no. 1810-154; *27* – sp. no. 4270-63.

14, 15 – Shapeless accumulations of coccoidal cells of *Myxococcoides* spp. (cells of *capsulata* and *punctata* degradation stages); *14* – sp. no. 4019-92, *15* – sp. no. 4271-60.

16 – *Nucellosphaeridium nordium* (Tim.), an envelope with distinct internal body; sp. no. 2378-137.

21 – Coenobial aggregate of small unisized envelopes of *Leiosphaeridia* sp.; sp. no. 2378-138.

22 – *Chuaria circularis* Walcott emend. Vidal, Ford; sp. no. 4233-62.

23, 24 – Filamentous forms with rounded sporangia-like structures consisting of small coccoidal cells located near the closed rounded termination of the filament; *23* – sp. no. 1810-155; *24* – sp. no. 1810-156.

25 – Branching filamentous form (sheath?); sp. no. 4453-41.

26 – *Entosphaeroides* sp., a fragment of filamentous sheath with numerous small dark internal bodies (collapsed cells?); sp. no. 4018-93.

28 – *Sphaerocongregus variabilis* Moor., chainlike aggregate of coccoidal cells without a sheath; sp. no. 4231-64.

30 – *Palaeopleurocapsa* sp.; aggregate of coccoidal cells with specific cluster-like structure; sp. no. 4233-66.

representatives of all recognized levels and their ecologic interpretation.

Stratigraphic and facial-ecological distribution of the Derevnya microfossils. The first level (the lower subformation; Nizhnyaya Tunguska locality downstream from the Strel'nye Gory, 59 - 61 m above the formation base) contains an impoverished assemblage of microfossils represented by acritarchs, collonial coccoidal and filamentous forms, and rare thalluses and ribbonlike structures with narrow, dark longitudinal zones. The assemblage is known from two samples separated by 3.3 m in the section. The stratigraphically lower sample is dominated by large complicated forms of the above morphotypes, whereas the upper one comprises smaller acritarchs and simple narrow sheaths. The first sample yielded the following: acritarchs *Leiominuscula*, *Leiosphaeridia*, and *Chuaria* (diameter up to 800 μm);[1] envelopes with internal bodies of *Nucellosphaeridium* (diameter up to 600 μm) and with peripheral zones *Simia* (the outer diameter from 30 to 110 μm); small aggregates of coccoidal *Myxococcoides* (cell diameters of 8 - 12 and 20 - 24 μm); smooth-walled sheaths of *Eomycetopsis*, *Leiotrichoides*, and *Asperatofilum* (2 - 4, 6 - 20, and 30 - 50 μm wide respectively); multiserial *Polytrichoides*; and a large number of large septate filamentous *Botuobia* (70 - 150 μm wide). Wide (50 - 100 μm) thalluses, including those with dark longitudinal zones (70 - 72 μm wide), also occur there. The second sample contains acritarchs *Leiominuscula* and *Leiosphaeridia* (10 - 110 μm), single specimens of *Myxococcoides* (cell sizes are 8 - 12 and 14 - 16 μm), abundant sheaths of *Eomycetopsis* and *Leiotrichoides* (2 - 24 μm wide), rare fragments of multiserial *Polytrichoides*, and trichomes of *Botuobia* (70 - 84 μm).

These taxa are derived from thin (1 - 6 cm) black shale interbeds. The lower interbed sandwiched between horizontally bedded dark-grey dolomites has a sharp lower boundary and grades upward steeply into the dolomites. The upper interbed is bounded by thick

[1] We assign to genus *Chuaria* Walk. emend. Vidal et Ford, 1985, the larger (more than 300 μm in diameter) sphaeromorphic acritarchs with dense acid-resistant envelopes with characteristic folds. We attach special evolutionary and taxonomic significance to this size limit, considering that the phytoplanktonic organisms exceeded it only at the beginning of the Late Riphean (Veis, 1988, 1993).

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Plate I.

Plate II. Microfossils of the Upper Riphean Derevnya Formation of the Turukhansk Uplift.

All specimens are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

Levels of the Derevnya microfossil occurrence:

14 – the first level; *1 - 6, 9, 11 - 13, 15 - 19, 23, 25, 31, 32* – the second level; both in the lower subformation (Nizhnyaya Tunguska River downstream of the Strel'nye Gory); *20, 29* – the third level, *24* – the fourth level; both in the middle subformation, the lower siliciclastic member, the same locality.

7, 8, 10, 21 – the sixth level; the middle subformation, the lower siliciclastic member (Miroedikha Vtoraya River).

30 – the eighth level; *22, 26 - 28* – the tenth level; *22, 26 - 28, 30* – eleventh level; all three in the middle subformation, the upper siliciclastic member (Nizhnyaya Tunguska River downstream Strel'nye Gory).

Bar for *1, 22* – 240 μm; for *2, 4* – 250 μm; for *3* – 200 μm; for *5 - 7, 9, 31* – 150 μm; for *8, 23, 27, 29, 32* – 100 μm; for *10, 21* – 350 μm; for *11 - 13, 15 - 20, 25, 26, 28* – 80 μm; for *14* – 400 μm; for *24, 30* – 120 μm.

1 - 4 – *Obruchevella cylindrica* (Tynni et Donner) emend. Jank., cylindric spirals composed of smooth-walled filamentous sheaths; *1* – sp. no. 4231-56; *2* – sp. no. 4231-67; *3* – sp. no. 4231-68; *4* – sp. no. 4231-69.

5, 6, 26 – *Glomovertella glomerata* (Jank.), hemilike aggregates of filamentous sheaths; *5* – sp. no. 1810-158; *6* – sp. no. 4233-70; *26* – sp. no. 2378-133.

7 – *G. eniseica* (Herm.), shapeless ball-like tufts of filamentous sheaths; sp. no. 4019-92.

8, 10, 22, 28, 30 – *Siphonophycus* sp., broad filamentous sheaths with distinct cross ornamentation; *8* – sp. no. 4018-95; *10* – sp. no. 4018-94; *22* – sp. no. 1882-40; *28* – sp. no. 2378-143; *30* – sp. no. 1880-41.

9 – *Rugosoopsis tenuis* Tim. et Herm., tubular costate form; sp. no. 4231-71.

11 - 13, 15 - 19, 23, 25 – *Archaeoellipsoides* (=*Brevitrichoides*) *grandis* Horod. et Donald., large Akinette-like ellipsoidal envelopes with enclosed trichome remnants; *11* – sp. no. 1810-159; *12* – sp. no. 1810-160; *13* – sp. no. 4231-72; *15* – sp. no. 1810-161; *16* – sp. no. 1810-162; *17* – sp. no. 1810-163; *18* – sp. no. 1810-164; *19* – sp. no. 1810-165; *23* – sp. no. 4231-73; *25* – sp. no. 1810-166.

20 – *Eosynechococcus* sp., small ellipsoidal envelopes without internal inclusions; sp. no. 2378-140.

21 – *Leiotrichoides typicus* Herm., a fragment of interlacing mat consisting of loosely spaced smooth-walled sheaths; sp. no. 4018-95.

24 – *Brevitrichoides tiulmenicus* Jank. et Kozl, irregular spindlelike envelope with supposed multichamber internal structure; sp. no. 1810-166.

27 – *Rectia costata* (Jank.), wrinkled-costate tubular form; sp. no. 2378-111.

29 – *Siphonophycus attenuatum* A. Weiss, tapering sheath with transverse narrow reduced zones; sp. no. 2378-144.

31 – *Leiotrichoides gracilis* Pjat.; sp. no. 4273-75.

32 – filamentous sheaths with diagonal narrow reduced zones; sp. no. 4270-76.

(up to 0.5 m) beds of coarse-grained cross-bedded wackestones with isolated bioherms of *Baicalia* (the clastic dolomite association of the second type). These deposits formed during the low sea-level stand at the initial stage of a transgression (Fig. 2). The clayey sediments of the lower interbed were deposited in the quiet-water environment of small shallow semi-isolated lagoons, which appeared at the earliest stages of flooding. In the northern part of the region, these sediments are replaced by fine clayey-carbonate deposits with dissolution marks (probably karst) and local erosional surfaces, especially characteristic of the Kamennaya River section (coastal semiisolated lagoons). On the contrary, the clayey matter of the upper interbed was accumulated in a small protected floor area related to the wave shade of small stromatolite bioherms. These sediments were buried under the storm-mobilized fine carbonates.

The second level (the lower subformation; Nizhnyaya Tunguska River downstream from the Strel'nye Gory, 71 - 72 m above the formation base) is represented by five samples from a 1-m-thick interval of the section. The microfossil assemblage of this level is outstanding because of its abundance and taxonomic diversity as compared to all other assemblages of the Derevnya Formation. Its specific feature is an abundance of various collonial coccoidal, filamentous and large ellipsoidal forms. The colonial coccoids can be subdivided into five morphological groupings: (1) small simple shapeless aggregates of *Myxococcoides* (the cell diameter in the colonies is from 8 to 30 μm); the cells of some colonies with an unusual reticular appearance have thick (up to 3 μm) walls; (2) large, variously shaped thalluses of *Ostiana*, composed of closely spaced coccoidal cells (including those of the *punctata* degradational stage; Hofmann, 1976) of several size generations (10, 18, and 30 - 40 μm); (3) very abundant chainlike colonies of *Sphaerocongregus* (probably *Polysphaeroides* and/or *Chlorogloeopsis*; Mikhailova, 1986, 1987; Yankauskas, 1989; Hofmann and Jackson, 1994), which constitute uni- to triserial aggregates, the uniserial ones with cell diameters of 10 and 14 μm, and the other with dimensions of 10 - 12 and 20 - 24 μm; (4) clustered colonies resembling *Pleurocapsa* (cell size 22 - 24 μm); and (5) rare solitary ellipsoidal *Eosynechococcus* (12 × 20 - 30 μm).

The sheaths and trichomes are present at the second level in approximately equal proportions. The sheaths include the following: simple *Archaeotrichion*, *Eomycetopsis*, *Leiotrichoides* and *Asperatofilum* (from 1 - 2 to 50 μm in width); multiserial *Polytrichoides* in association with rare wrinkled-costate *Rectia* (up to 60 - 80 μm wide); ring-shaped and shapeless tufts of smooth-walled *Glomovertella* (filament width of 3 - 10 μm); and cylindric spirals of *Obruchevella* (ring diameter 60 μm, filament width 6 - 8 μm). The trichomes comprise typical *Oscillatoriopsis* (6 - 8 μm wide), filaments

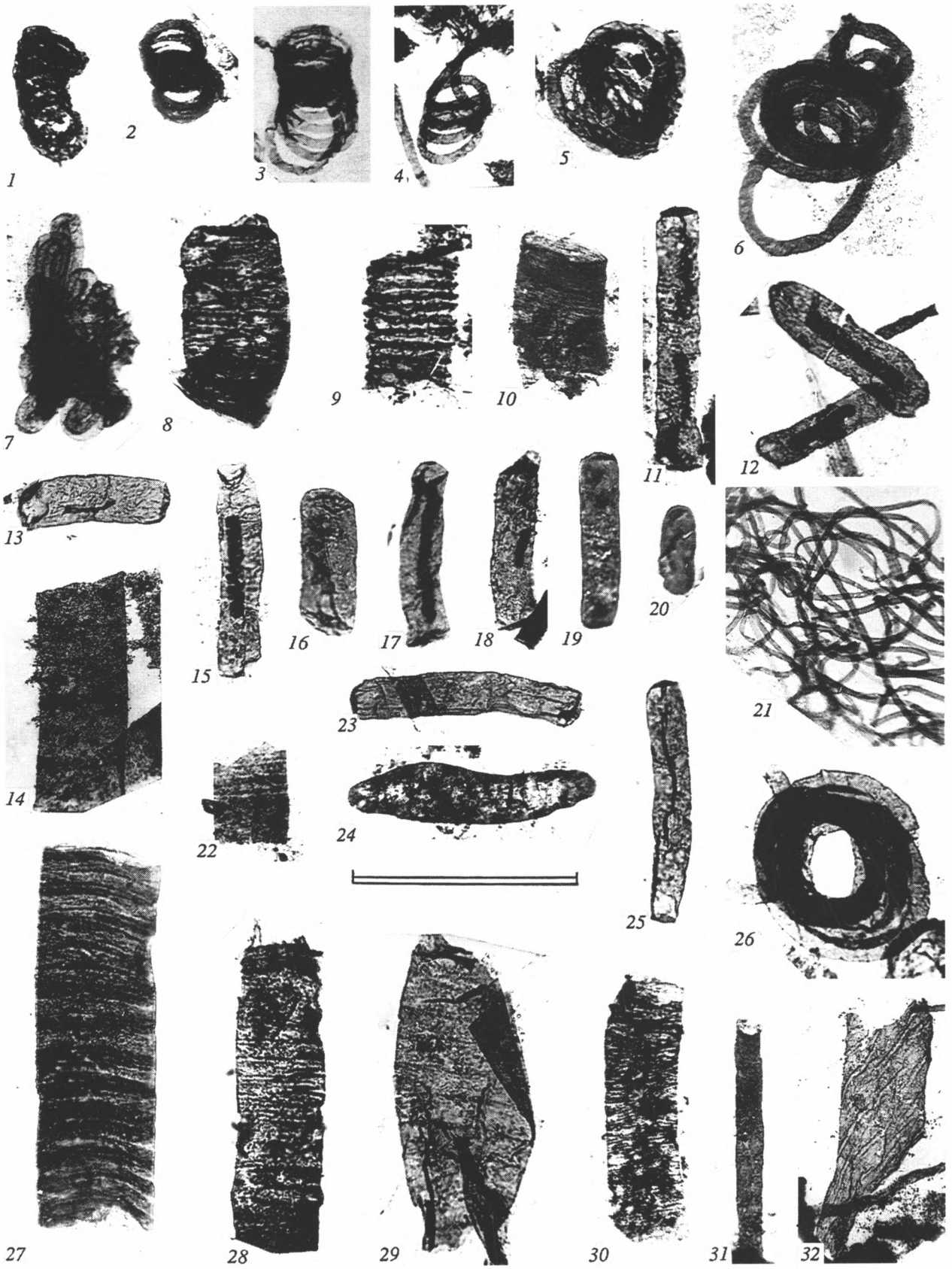
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Plate II.

Plate III. Microfossils of the Upper Riphean Derevnya Formation of the Turukhansk Uplift.

All specimens are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

Levels of the Derevnya microfossil occurrence:

1 - 3 – the first level; *4, 5, 7, 9, 11, 12, 16, 20 - 22, 25 30* – the second level; both in the lower subformation (Nizhnyaya Tunguska River downstream of the Strel'nye Gory);

8, 14, 19 – the third level; the middle subformation, the lower siliciclastic member, the same location.

6, 15, 18, 24 – The sixth level, the middle subformation, the lower siliciclastic member (Miroedikha Vtoraya River).

13 – The ninth level; the middle subformation; the upper siliciclastic member (Kamennaya River).

10, 17, 23 – The tenth level; the middle subformation, the upper siliciclastic member (Nizhnyaya Tunguska River downstream Strel'nye Gory).

Bar for *1, 16* – 300 μm; for *2, 14, 28* – 550 μm; for *3, 18, 24* – 400 μm; for *4, 9, 17, 20* – 120 μm; for *5, 12, 13, 22, 29* – 150 μm; for *6, 11, 21* – 80 μm; for *7, 8, 10, 15, 19, 23, 26, 30* – 100 μm; for *25, 27* – 200 μm.

1, 2 – *Botuobia patomica* Kolos., large trichomelike structures; *1* – sp. no. 1800-23; *2* – sp. no. 1800-24.

3 – *B. magna* (Tynni et Donner) emend. Kolos., medium-sized trichome-like structures; sp. no. 1800-25.

4 – A filamentous remnant (trichome?) ornamented transversely and consisting of elongate constricted sections (ex gr. *Peleduica*); sp. no. 4271-77.

5, 20 – *Oscillatoriopsis* sp., trichomes with elongate cells and distinct septs; *5* – sp. no. 4271-78; *20* – sp. no. 4231-82.

6 – *Palaeolyngbya catenata* Herm., trichome remnants in distinct filamentous sheath; sp. no. 4018-96.

7 – *Leiotrichoides gracilis* Pjat., cylindric sheath with closed, rounded and slightly constricted termination; sp. no. 1810-167.

8 - 11, 17, 22, 23 – *Oscillatoriopsis media* Mend. et Schopf; *8* – sp. no. 2378-145; *9* – sp. no. 4271-79; *10* – sp. no. 2378-146; *11* – sp. no. 1810-168; *17* – sp. no. 2378-148; *22* – sp. no. 1810-170; *23* – sp. no. 2378 -149.

12 – A fragment of *Stiganema*-like filamentous remnant; sp. no. 4233-80.

13 – *Trachytrichoides ovalis* Herm., lenticular trichomelike structures; sp. no. 4453-50.

14 – *Taenitrichoides jaryshevicus* Asseeva, very broad filamentous forms (sheaths?); sp. no. 2378-147.

15 – *Calyptothrix* sp., specific crenulated trichomes; sp. no. 4015-97.

16 – A form of unusual structure and irregular ellipsoidal outlines; sp. no. 4231-190.

18 – *Plicatidium latum* Jank., broad filamentous structures with transverse indistinct ornamentation; sp. no. 4018-98.

19 – An elongate remnant with dark internal zone; sp. no. 2378-148.

21 – *Oscillatoria*-like trichome with damaged cellular structure; sp. no. 1810-169.

24 – Cuticular covers with coarse transverse ornamentation, which belong, supposedly, to the organisms of animal origin, having coarse transverse ornamentation; sp. no. 4018-99.

25 – Ribbonlike structureless thalluses with distinct intercalar swell-bodies (ex gr. *Archaeoclada*); sp. no. 4231-83.

26 – Chains of large ellipsoidal envelopes of different size with enclosed trichomes; sp. no. 4233-84.

27 – Germinated Fungi-like forms ex gr. *Germinosphaera unispinosa* Mikh.; sp. no. 4233-85.

28 – Large short ellipsoidal envelope with dark granulated content filling in its inner chamber (ex gr. *Fabiformis*); sp. no. 1810-178.

29 – *Polytrichoidea lineatus* Herm., aggregate of parallel filamentous sheaths; sp. no. 1810-171.

30 – *Majasphaeridium carpogenum* Herm., a fragment of thallus with distinct bilobed structure; sp. no. 2378-150.

of *Palaeolyngbya* (the trichomes are 10 - 14 μm wide and the sheaths are 16 - 20 μm wide), and some other small forms (5 or 6 μm wide). In addition to common cells, the latter forms include larger (up to 16 - 20 μm) elongate structures similar to heterocysts and/or spores of some present hormogonian cyanobacteria. Particularly noteworthy elements of the second-level assemblage are tubular *Entosphaeroides* (diameters 18 and 22 - 26 μm), retaining closed rounded terminations and including small (2 - 6 μm) coccoidal cells inside. At this level, there occur earlier undescribed branching filamentous *Stiganema*-like remnants and their fragments, representing thin-walled cylindric sheaths (30 - 32 μm wide) with internal uni- and biserial chainlike aggregates of cells, 8 - 30 μm in diameter. These filamentous forms constitute a single morphological grouping with large elongate vesicles of *Archaeoellipsoides* (=*Brevitrichoides*),[2] ranging in size from 15 - 20 × 50 - 60 to 10 - 20 × 100 (? 280) μm and showing ellipsoidal to subrectangular shapes. Many of these vesicles with dark elongate internal bodies (trichomes) are likely to be well-known Precambrian hormospores of some filamentous *Nostoc*- and *Stiganema*-like cyanobacteria. Other similar but shorter forms (partly with internal inclusions) compose chains of variable length, and show a close similarity with *Arctacellularia* and *Trachytrichoides*, the remnants of uncertain taxonomic affinity (probably green algae).

The acritarchs of the second level are less abundant and diverse. The simplest *Leiosphaeridia* (diameter up to 150 μm) and their cenobial aggregates only occur in relative abundance. Much less frequent forms are *Chuaria* (up to 500 μm), envelopes of *Leiosphaeridia bicrura* Jank. with medium splits, and *L. ternata* (Tim.) with broken peripheral zones as well as *Pterospermopsimorpha*, *Nucellosphaeridium*, *Simia* of similar size (up to 160 - 170 μm) and single forms of ?*Leiofusidium* with two winglike processes (diameter 40 μm). The complex microfossils include spindle-shaped vesicles of supposed chamber structure, specific dumbbell-

[2] In previous publications (Veis and Petrov, 1994a, 1994b; and others) such forms were referred to genus *Eosynechococcus* Hofmann.

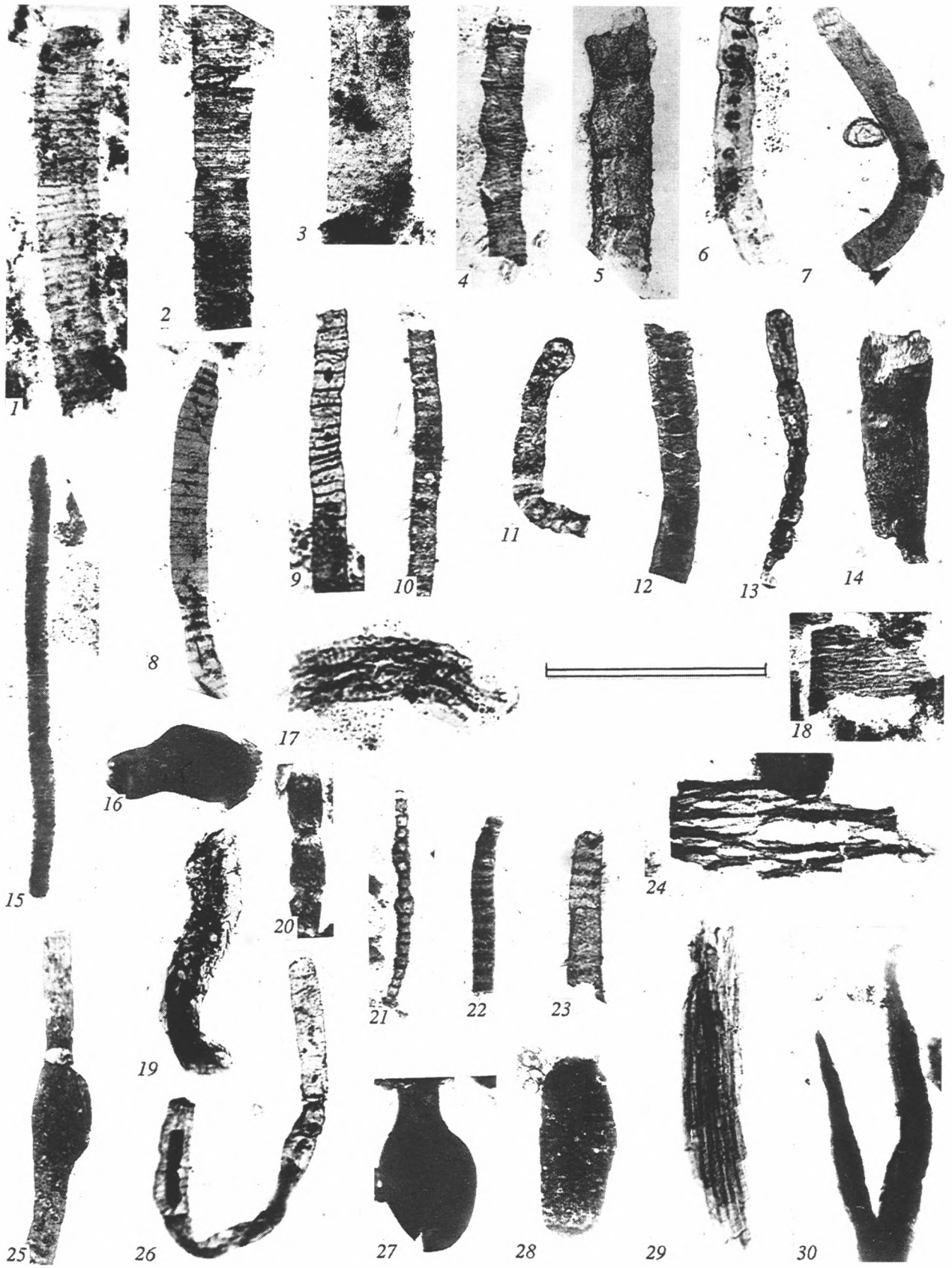
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Plate III.

shaped forms, broad (up to 100 μm) dense thalluses with a longitudinal structure, and rare envelopes with single short or long processes related to *Germinosphaera* (envelope diameter from 14 to 28 - 30 μm, process length 5 or 6 μm) and *Caudosphaera* (envelope diameter up to 140, process length more than 100 μm). The two latter taxa are commonly considered to be similar to the germinating structures of some eumycete fungus (German, 1979, 1990; Burzin, 1992).

The peculiar features of this assemblage may be attributed to the specific environment of its formation. The assemblage occurs in a dark-grey and greenish-grey shale lens, 1 m thick, inside clayey-carbonate deposits filling a narrow (5 - 10 μm) and deep (up to 10 m) interbiohermal depression. The stromatolite bioherms of *Jacutophyton* and *Baicalia* and enclosing sediments were formed at a low sea-level stand in a relatively deep-water environment (below the permanently active wave base). At the time of accumulation of the clayey sediments enclosing the described assemblage, the synoptic relief of the depression probably did not exceed a few meters, and the bioherm tops reached the above base. Later, this relief was leveled and overlain by biostromes of columnar and columnar-layered stromatolites, which were formed in relatively shallow settings with permanent wave activity.

The third level (the middle subformation; the lower siliciclastic member, Nizhnyaya Tunguska River downstream from the Strel'nye Gory, 110 - 112 m above the formation base) is represented by four samples from a 2-m interval of the section. The microfossil assemblage shows nearly equal proportions of members of the main morphological groupings, a great number of specimens, and a lack of the largest morphotypes. Acritarchs are dominated by *Leiosphaeridia* (up to 150 μm in diameter), including *L. bicrura* Jank. and *L. ternata* (Tim.), but larger *Leiosphaeridia* and *Chuaria* are rare. Associated forms are *Pterospermopsimorpha* (diameter 30 - 80 μm) and the first *Trachyhystrichosphaera* of the Derevnya Formation (envelope fragments are about 50 μm in size). Of colonial coccoidal forms, small aggregates of *Myxococcoides* (cell diameter 8 - 24 μm) only occur abundantly in association with less frequent thalluses of *Ostiana* (cell diameter up to 10 μm) of the *capsulata* and *punctata* degradation stages, biserial chainlike *Sphaerocongregus* (cell diameter 20 - 24 μm), some structures comparable with *Palaeopleurocapsa* (cell diameter 10 - 14 μm), and ellipsoidal *Eosynechococcus* (4 - 5 × 20 - 22 μm).

Among filamentous microfossils, the following forms were identified: smooth-walled sheaths of *Eomycetopsis*, *Leiotrichoides*, *Asperatofilum*, and *Taenitrichoides* showing a continuous dimensional range from 2 to 90 μm; cylindric crenulated *Rugosoopsis* (44 - 46 μm wide); abundant wrinkled-costate *Rectia*, which are exceptionally well preserved; multiserial *Polytrichoides*; tapering *Siphonophycus attenuatum* A. Weiss; and elongate *Archaeoellipsoides* (= *Brevitrichoides*) (14 - 16 × 100 - 110 μm), including those with a coccoidal cell attached to one of the terminations. Similar forms were previously recorded in the Kotuikan biota of the Anabar region (Sergeev, 1993). The trichomes of the third level are represented by unexpressive *Oscillatoriopsis* (18 - 20 μm wide) and *Botuobia* (up to 70 μm). The assemblage also comprises filamentous remnants of two-dimensional generations with dark longitudinal zones (13 to 14 and 25 μm in width).

The microfossil assemblage of the third level is confined to the sandy-clayey rhythmite facies of the lower horizons of the shallow-water complex (Fig. 3). The rock deposition occurred in the proximal environments of the upper subtidal zone of the inner shelf near the coastal mobile sandy belts. The progradation of these belts determined the regressive shallowing trend that is characteristic of this time interval.

The fourth level (the middle subformation; the lower siliciclastic member, Nizhnyaya Tunguska, 124 - 127 m above the formation base) is represented by eight microfossiliferous samples derived from a uniform clayey-silty member, about 3-m thick (Fig. 3). The lower part of the level contains only the following: small acritarchs *Oleiosphaeridia* (10 - 30 m, less frequently up to 50 μm), accompanied by single and similarly sized *Nucellogphaeridium*; the simplest *Myxococcoides* (cell diameter in colonies 14 - 30 μm); rare *Eomycetopsis*, *Leiotrichoides* (2 - 20 μm wide); and *Palaeolyngbya* (trichome from 10 to 14 and sheath up to 20-μm wide); and branching thalluses of *Majaphyton* and *Ulophyton*. The Derevnya Formation displays these thalluses in this assemblage only.

In the upper part of the fourth level, the microfossils increase greatly both in diversity and maximal size limits. In addition to relatively simple *Leiosphaeridia* and *Chuaria* (up to 700 μm in size), marginated *Simia* (diameter 110 - 120 μm), and envelopes with winglike processes (?*Leiofusidium*), there appear acanthomorphic acritarchs *Trachyhystrichosphaera aimica* Herm. (diameter 200 - 210 μm) and *Prolatoforma aculeata* Mikh. (290 × 360 μm in size). The less frequent colonial coccoidal forms include fragments of *Myxococcoides* colonies (cell diameter 8 - 14 μm). Their infrequent occurrence is more than compensated for by the abundance and diversity of filamentous microfossils. In addition to the representatives of this grouping known from the lower part of this level, the following occur: broad (up to 100 μm) smooth-walled *Asperatofilum* and *Taenitrichoides*; poorly wrinkled *Plicatidium* (200 μm and more in width); coarse-costate *Rectia*; sheaths with closed terminations, undistinguishable from *Entosphaeroides* but deprived of cellular content; and some other forms. As for trichomes, the upper part of the level yielded only *Oscillatoriopsis*, 12 μm wide, with intercellular spots. Complex structures are represented by ribbonlike dark thalluses, 50 and 100 μm wide, and sporadic sheaths similar to *Brevitrichoides tjulmenicus* Jank. et Kozl., which are 125 × 65 - 70 μm

in size and have an irregular spindlelike shape and, probably, a multichamber structure.

The sediments enclosing the microfossils of the fourth level were deposited during the initial stage of formation of the transgressive complex of the lower siliciclastic member (Fig. 3) in the course of rapid flooding of the shallow flat. The increase in microfossil diversity from the base to the top of the fourth level correlates with the depth growth. This is emphasized by the transition form siltstones to shales in this interval of the section.

The fifth and sixth levels (the middle subformation; the lower siliciclastic member, Miroedikha Vtoraya River, 35 - 45 m above the visible formation base) are derived from the reduced section of the lower siliciclastic member of the eastern monocline (Fig. 3), which comprises two sandy-clayey packages. Six samples of the fifth level are from the lower package, which is 1 m thick and represented by siltstone facies. The characteristic features of this interval are abundant local erosional surfaces, fine-grained sandstone interbeds, medium-grained sandstone molds of relatively narrow (up to 15 cm) meandering erosional channels (gutter casts), and a gradually decreasing amount of sandy matter upward in the section. The package overlies with a sharp boundary the coarse-grained sandstones with carbonate cement, the lowermost exposed unit of the siliciclastic member. Seven samples of the sixth level come from the upper package, which is 3 m thick and represented mainly by shale facies. The base of the package displays a number of small local erosional surfaces; the silt content rapidly decreases in abundance from here upward, but glauconite grains are abundant. At the top of the package, the shales grade into the clayey-carbonate deposits (Fig. 3).

Microfossils of both levels are dominated by relatively large acritarchs *Leiosphaeridia* (diameter from 30 to 260 - 280 μm); sheaths of *Eomycetopsis* and *Leiosphaeridia* (2 - 20-μm wide), which often form compactly interwoven mats; ball-like *Flomovertella*; multiserial *Polytrichoides*; and ribbonlike nonbranching thalluses (40 - 80 μm wide). Broad sheaths of *Asperatophilum* (diameter up to 50 - 70 μm) and trichomes of *Botuobia patomica* Kozlov (36 - 40-μm wide) are less frequent. In addition, the fifth level contains acritarchs *Nucellosphaeridium* (diameter 120 μm), *Leiosphaeridia bicrura* Jank., short-spined *Trachyhystrichosphaera aimica* Tim. et Herm., smooth-walled sheaths of *Taenitrichoides* (70 μm and more in width), trichomes of *Oscillatotiopsis* (6 - 8 μm wide), ellipsoidal vesicles of *Archaeoellipsoides* (18 × 100 μm), trichomelike *Trachytrichoides*, and dark structureless thalluses (up to 100 μm wide).

The taxa specific to the sixth level only include the following: acanthomorphic acritarchs *Trachyhystrichosphaera stricta* Germann (diameter up to 440 μm) and *Prolatoforma* (up to 500 μm), forms with internal bodies such as *Nucellosphaeridium* (up to 450 μm) and *Pterospermopsimorpha* (up to 430 μm), cenobial aggregates of *Leiosphaeridia* (vesicle diameter up to 70 μm), thalluses of coccoidal *Ostiana* with thin- and thick-walled cells, abundant broad sheaths of ?*Siphonophycus* with specific transverse wrinkles (up to 50 and even 76-μm wide) or light reduced zones (70 - 80 μm wide), very large ribbonlike *Plicatidium* (some of them up to 360 μm wide), and diverse *Oscillatoria*-like trichomes (up to 30 μm wide). Some unique elements of the Derevnya microbiota, namely coarse-costate cuticular remnants (tube width 80 and 160 μm) and especially broad (up to 200 μm) ribbonlike thalluses of longitudinal structure, were also found.

The sandy-clayey packages containing microfossils of the fifth and sixth levels compose a single transgressive succession of two deepening trends (Fig. 3). The boundary between these trends, located at the base of the upper package, is considered to be a conformable transgressive surface. Thus, the sediments of the fifth level were accumulated in shallow upper subtidal settings, whereas those of the sixth level were accumulated in relatively more distal and deeper areas of the lower subtidal zone. The most distal shelf area was predominantly featured by carbonate sedimentation synchronous to the clayey deposition of the fifth level (Fig. 4).

The seventh level (the middle subformation, the upper siliciclastic member, Nizhnyaya Tunguska River downstream from the Strel'nye Gory, 177 - 180 m above the subformation base) is represented by six samples from the upper part of the sandy-clayey package, 12 m thick. This level yielded a moderately diverse microfossil assemblage with abundant acritarchs; other morphotypes are rare. Upsection, this assemblage becomes somewhat impoverished, except for acritarchs dominated by small *Leiosphaeridia* (diameter 20 - 50 μm) in association with rare larger envelopes of this genus and sporadic *Chuaria* (400 - 500 μm in size). In addition, *Nucellogphaeridium* forms (50 - 100 μm) were found in the lower part of the interval, and those of *L. ternata* (Tim.) (up to 400 μm) and *L. bicrura* Jank. were found in the upper part. As for coccoidal forms, the lower part contains *Myxococcoides* (cell diameters in the colonies are 14 - 18 and 26 - 30 μm), dense sphaeroidal accumulations of small cells (up to 5 to 6 μm) of *Palaeoanacystis*, and sporadic colonies similar to palmelloid *Eoentophysalis* and pseudofilamentous *Palaeopleurocapsa* (cell sizes 5 to 6 and 10 - 14 μm, respectively). The upper part includes only the simplest forms of *Myxococcoides* (cell diameter 14 - 16 μm). Filamentous microfossils of the lower part of the level are represented by abundant sheaths of *Eomycetopsis*, *Leiotrichoides*, and *Asperatofilum* (from 2 to 66 μm in width) and complex forms associated with structureless ribbonlike thalluses (50 - 60 μm wide) and fungilike *Caudosphaera* and *Germinosphaera* (vesicle size 60 - 160 μm). Upward, the complex forms disappear completely, and sheaths are repre-

sented only rarely *Eomycetopsis* and *Leiotrichoides* (6 - 12 μm).

The listed microfossils are enclosed in the siltstone facies of the regressive complex, corresponding to the continuous shallowing-up succession, which shows a noticeable increase in the grain size of the material upward in the section. These deposits were formed in the proximal lower subtidal settings against the background of the rapid progradation of siliciclastic masses in the vicinity of advancing sandy belts of the upper peritidal zone (Fig. 3).

The eighth level (the middle subformation, the upper siliciclastic member; Nizhnyaya Tunguska River downstream from the Strel'nye Gory, 170 - 175 m above the formation base) is represented by two samples from a thin (about 2 m) package of frequently alternating shales, siltstones, and sandstones. The level contains acritarchs *Leiosphaeridia* and *Chuaria* (diameter from 20 - 30 to 400 μm), rare *Nucellosphaeridium* (diameter up to 100 - 200 μm), and single acanthomorphic forms *Trachyhystrichosphaera truncata* Hermann et Jank. with very short spinelike processes (vesicle diameter 160 μm). Coccoidal microfossils are represented only by *Myxococcoides*, which constitute shapeless accumulations (cell size 26 - 30 μm) or oval colonies of smaller cells (14 - 18 μm). A set of filamentous forms consists of smooth-walled sheaths of *Eomycetopsis*, *Leiotrichoides*, *Asperatofilum*, and *Taenitrichoides*, ranging in width from 2 to 88 μm (usually 20 and 54 - 56 μm) and associated with wrinkled-costate *Rectia* (44 - 46 μm wide). Complex structures were not found. The deposits of this level are included in the basal beds of the shallow-water complex of the upper siliciclastic member (the sandy-clayey rhythmite facies), formed in the upper subtidal environments of the distal part of the shelf.

The ninth level (the middle subformation, the upper siliciclastic member, Kamennaya River, 12 - 15 m above the visible member base) yielded two samples from a thin (3 m) sandy-clayey package. Its microfossil assemblage is distinguished by a very poor set of acritarchs, represented by rare *Leiosphaeridia* (diameter 70 - 140 μm), *Nucellosphaeridium* (diameter 120 μm), single oval envelopes (150 - 200 μm in size), and cenobial aggregates of smooth-walled envelopes (diameters 40, 50 - 60, and 80 μm) with or without internal inclusions. Colonial coccoidal and filamentous forms are much more diverse. The former consist of abundant *Ostiana* (cell sizes 10 - 12, 20 - 24, and 32 - 40 μm) and bi- or triserial chainlike *Sphaerocongregus* (cell diameter 8 - 10 and 18 - 20 μm). The sheaths are dominated by simple *Eomycetopsis* and *Leiotrichoides* (width from 2 to 30 μm), shapeless tufts of *Glomovertella* (filaments up to 10 μm), multiserial *Polytrichoides*, and specimens of with internal bodies *Entosphaeroides* (16 and 20 - 24 μm wide). Trichomes are represented by large *Oscillatoriopsis* (up to 24 μm wide) and chainlike aggregates of ex gr. *Trachytrichoides*, the cells of which, noticeably reduced in length from one termination to another (width 20 μm, length from 20 to 70 μm). They are also represented by small fragments of thin-walled sheaths with distinct lateral branching. The distinct elements of the assemblage are elongate ellipsoidal vesicles ranging from the smallest *Eosynechococcus* (12 × 30 μm) to much larger *Archaeoellipsoides* (= *Brevitrichoides*) and *Arctacellularia* (20 × 60, 14 × 50, and 16 × 70 μm) with dark trichomelike inclusions. The microfossils of the ninth level, as those of the eighth one, occur in the sandy-clayey rhythmite facies of the shallow-water complex of the upper siliciclastic member of the middle subformation.

The tenth level (the middle subformation; the upper siliciclastic member of the Nizhnyaya Tunguska locality, 194 - 197 m above the formation base) was especially thoroughly studied. Thirteen samples were collected from a 3-m-thick package of alternating siltstones and shales. Twelve samples yielded abundant and diverse microfossils of good preservation. One sample from the base of the package contains only a large amount of brown decomposed organic matter.

The assemblage of this level is very rich in acritarchs and filamentous fossils. The latter include sheaths and septate forms that are nearly equal in their proportions. The set of acritarchs consists of the following: simple *Leiominuscula* and *Leiosphaeridia* (including *L. bicrura* and *L. ternata*), *Chuaria* (up to 800 μm), *Nucellosphaeridium* envelopes with internal bodies (up to 120 μm), *Pterospermopsimorpha* (70 - 80 μm), *Konderia* (up to 220 μm along the long axis of the outer envelope), and rare prismatic *Octaedrixium* (20 μm). A general prevalence of envelopes of 70 μm in diameter is combined with an increase upward in the section in the abundance of forms with maximal dimensions.

The filamentous microfossils of the tenth level display a reverse trend in dimension variations. They are represented by the following: smooth-walled sheaths of *Eomycetopsis*, *Leiotrichoides*, *Asperatoofilum*, and *Taenitrichoides* (width from 2 to 90 - 120 μm); crenulated *Rugosoopsis* (37 - 56 μm wide); abundant wrinkled-costate *Rectia*; multiserial *Polytrichoides*; less frequent sheaths with closed terminations (ex gr. *Entosphaeroides* deprived of fragmental cellular content), ball-like *Tortunema* and *Glomovertella* (7 to 8 and 10 μm wide); and trichomes of *Cephalophytarion* (5 to 6 μm) and *Eotuobia* (34 - 90 μm). The lower third of this interval shows more diverse and particularly large specimens of filamentous forms. The largest specimens disappear in the middle part, where almost all *Rectia*, *Polytrichoides*, *Rugosoopsis*, *Tortunema*, *Glomovertella*, and *Entosphaeroides* were found. The diversity of the filamentous forms decreases sharply in the upper part of the interval until their complete disappearance near the top, where only single *Eomycetopsis*, *Leiotrichoides*, and *Asperatofilum* were recorded. Unlike acritarchs and filamentous forms, the coccoidal microfossils of the tenth level are extremely scarce and include

only *Myxococcoides* and, probably, *Palaeopleurocapsa* (cell size in colonies is 8 - 16 μm). They show a regular distribution throughout the section, being more frequent at the very top of the tenth level. Complex structures, all confined to the lower and middle parts of the interval, are represented by rare dark thalluses (50 - 100 μm), fungilike *Germinosphaera* (vesicle size up to 30 μm), and ribbonlike forms with longitudinal spots (24 - 40 μm wide).

This assemblage is confined to the shale facies of the transgressive complex of the upper siliciclastic member. The structure of the complex implies that it was formed during the progressive deepening of the paleobasin that affected the lower subtidal zone (the lower part of the tenth level) and the distal shelf areas (the upper part of the level).

The eleventh level (the middle subformation, Nizhnyaya Tunguska River downstream from the Strel'nye Gory, the upper siliciclastic member, 203 m above the formation base) is represented by a single sample containing a very impoverished microfossil assemblage. In addition to common *Leiosphaeridia* and *Chuaria* (diameters up to 300 and 800 μm, respectively), its acritarchs include a single specimen of a new form consisting of outer and inner envelopes, the latter locally covered with closely spaced thin processes (the outer envelope size is 520 μm). Other morphotypes belong to the following: coccoidal *Myxococcoides* (cell diameter 14 - 18 μm); sheaths of *Eomycetopsis*, *Leiotrichoides*, and *Asperatofilum* (width from 2 to 70 μm); and rare *Oscillatoriopsis* (6, 12, and 14 μm wide). This assemblage was found in the upper horizons of the transgressive complex. These horizons, corresponding to the mature stage of the complex development, were deposited in the deepest settings of the distal part of the Derevnya inner shelf, where combined clayey-carbonate sedimentation prevailed. Thus, this assemblage is related to the most distal and, probably, deepest as compared to deposits all these described above.

DISCUSSION AND MAIN CONCLUSIONS

The presented data characterize only time- and space-dismembered fragments of both the Derevnya biota and the relevant depositional settings. These fragments can be easily integrated in a single model of microfossil distribution across the Derevnya shelf. This integration provides new support for our previous conclusions that the lateral structure of the Riphean microfossil assemblages was under the predominant control of bathymetry (Veis and Petrov, 1994a, 1994b), and it allows us to distinguish for the first time some relationships between the assemblage diversity and a number of different abiotic parameters.

The small-scale parameters include the localization of an individual assemblage in a definite shelf zone, the confinement of assemblages to some specific elements of the basinal floor relief, and the hydrodynamic activity in the areas of their distribution. The main parameter is the microbiota relation to either the transgressive and regressive stages or the relatively stable stage of the basin development. The obtained data permitted an evaluation of the effect produced by each factor (including such important ones as basin depth) on the distribution of taxonomically different organic-walled microfossil assemblages and the dimensional generations of some characteristic morphotypes.

The facial-ecological zones[3] of particular Derevnya microfossil assemblages were spread throughout the relatively shallow shelf (from its distal area to the middle part of the upper subtidal zone), being populated very irregularly. The richest assemblages were concentrated in the distal (levels 4, 6, and 10) and proximal (levels 5 and 7 - 9) parts of the lower subtidal zone. The impoverished assemblages were confined to the middle part of this zone (levels 2 and, partly, 3), and the poorest biota occupied the deepest and shallowest zones of the Derevnya shelf (levels 11 and 1, respectively).

In addition to the specific localization of particular microbiotas, some general trends in their distribution can be seen. Most of the microbiotas evolved in partly isolated settings of the inner shelf with siliciclastic sedimentation. The open-sea settings with clayey-carbonate sedimentation appeared to provide little information with respect to microphytology. As for the deepest areas of the Derevnya basin with purely carbonate sedimentation (the rhythmical *Conophyton–Jacutophyton* biostromes) and the extremely shallow zones with coarse-grained siliciclastic sedimentation (the sandstone and clayey sandstone facies), these appeared, as was expected, to be free of organic-walled microfossils.

The bulk of the largest and complex Derevnya microfossils mainly inhabited the medium-depth interval corresponding to the lower subtidal zone and the distal limit of the upper subtidal zone (Fig. 5). All acanthomorphic acritarchs *Trachyhystrichosphaera* and *Prolatoforma*; nearly all representatives of such characteristic taxa as *Ostiana*, *Tortunema*, *Glomovertella*, *Obruchevella*, *Plicatidium*, *Archaeoellipsoides*, *Germinosphaera*, and *Caudosphaera*; broadest *Taenitrichoides*; and other noticeable microfossils were distinguished in this interval. As to the assemblages from the deeper and shallower shelf areas, they are much less diverse.

This morphotype distribution pattern could be interpreted as supporting the earlier conclusions on the existence of several facial-ecological fossil groupings in the Riphean basins (Veis and Petrov, 1994a, 1994b). The medium-depth assemblages of levels 2 - 10 could be assigned to the identified "optimal" grouping, and microfossils of levels 11 and 1, to the deep-water and

[3] Later, it would be useful to distinguish the wider "potential" and narrower "realized" facial-ecological niches along the basin profile. The latter correspond to the paleoenvironments where an assemblage was found, and the former, to the entire distribution of the relevant facies in the paleobasin.

shallow-water (or transitional) groupings, respectively. However, three obvious facts prevent such an interpretation. First, microbiotas of levels 1 and 11 contain several forms characteristic of the "optimal" grouping of other regions and age intervals. These are the following: particularly large *Chuaria*, *Simia*, *Rectia*, *Polytrichoides*, broad *Botuobia*, and ribbonlike thalluses found in the first level of the Derevnya Formation, as well as vesicles of unusual structure with processes and large smooth-walled sheaths from the eleventh level, etc. Second, most of the typical components (both planktonic and benthic) of the "optimal" grouping of the Derevnya shelf originated in shallower water environments than comparable forms of other regions. Third, the proximal and distal parts of the Derevnya shelf do not show a replacement of the "optimal" grouping by specific shallow-water, transitional, or deep-water groupings. The shallow-water limit of the "optimal grouping" is unknown in the Derevnya shelf, whereas toward the deep-water zone, this grouping shows a rapid and nonselective reduction in its composition. Farther off shore, at the transition to the deepest association of rhythmical biostromes, microphytological information is completely lost.

The above suggests two alternatives: either the specific local conditions of the Derevnya basin allowed some members of the "optimal" grouping to penetrate into unusual paleoenvironments, or the revealed lateral distribution of the Precambrian biotas proves to be correct only for three of the regions studied earlier. From what is given below, we can see the validity of the first alternative.

The microfossil distribution in relatively deep-water paleoenvironments is the easiest to explain. The sharp decrease in the assemblage diversity characterizing the transition from the distal (levels 4, 6 and 10) to the open-sea part of the shelf (level 11) seems to be very significant. This fact and some typical elements of the "optimal" grouping, which occur in the relatively deep-water assemblage of the eleventh level, suggest that the material under consideration lacks not only the deep-water facial-ecological grouping, but the distal assemblages of the "optimal" zones as well. This lack might be caused by the unusually wide extension of the "optimal" grouping in the very gentle Derevnya shelf. The microphytological record from this area is far from complete and, as noted above, terminates with depth growth because of the rapid replacement of clayey sediments by carbonates free of microfossils.

It is more difficult to explain the penetration of some members of the Derevnya "optimal" grouping (large and/or complex planktonic and benthic morphotypes) into the relatively shallow shelf zone, i.e., into the proximal area of the lower subtidal zone and in the upper subtidal zone. It should be stressed that the mentioned forms were concentrated in the outer open shelf with carbonate sedimentation, adjacent to the permanently active wave base (level 1), whereas their concentration in the inner protected siliciclastic shelf (levels 3, 5, and 7 - 9) tended to be confined to the upper subtidal zone, the area of weakening tidal currents.

The above suggests that hydrodynamic activity, just as paleobasin depth, served as a barrier against the "optimal" grouping penetrating into paleoenvironments

Fig. 5. Micropaleontological characteristics of section levels and their bathymetric and facial position in the Derevnya shelf.

a - c – Trends of the basin development (the dynamics of sea-level changes): a – transgressive, b – regressive, c – relatively stable; figures in circles denote numbers of the levels with microfossils (for other symbols see Fig. 3). Environments: d. sh. – distal shelf; l. sl. – a lower sublittoral zone; u. sl. – an upper sublittoral zone; l. – a littoral zone. *1 - 68* – Organic-walled microfossils: CC – colonial coccoidal microfossils; CM – complex microfossils.

1 – *Leiominuscula minuta* Naum.; *2* – *Leiosphaeridia* spp.; *3* – *Chuaria circularis* Walc. emend. Vidal, Ford; *4* – *Nucellosphaeridium nordium* (Tim.) and/or *N. minutum* Tim.; *5* – *Simia annulare* (Tim.) emend. Mikh.; *6* – *S. nerjuenica* A. Weiss; *7* – *Leiosphaeridia bicrura* Jank.; *8* – *Pterospermopsimorpha pileiformis* Tim. emend. Mikh.; *9* – *P. capsulata* Jank.; *10* – *Leiosphaeridia ternata* (Tim.) emend. Mikh. et Jank.; *11* – *Leiofusidium* sp.?; *12* – *Konderia elliptica* A. Weiss; *13* – *Octaedrixium* sp.; *14* – *Prolatoforma aculeata* Mikh.; *15* – *Trachyhystrichosphaera aimica* Herm.; *16* – *T. stricta* Herm.; *17* – *T. truncata* Herm. et Jank.; *18* – acanthomorphic envelopes with closely spaced thin processes in outer membrane; *19* – large acritarch-like ellipsoidal envelopes; *20* – *Palaeoanacistis vulgaris* Schopf; *21* – *Myxococcoides minor* Schopf; *22* – *M. reticulata* Schopf and/or *M. inornata* Schopf; *23* – *M. grandis* Nor. et Donaldson; *24* – *Ostiana microcyctis* Herm.; *25* – *Palaeopleurocapsa* sp.; *26* – *Eoentophysalis* sp.; *27* – *Sphaerocongregus variabilis* Moor.; *28* – *Conjunctiophycus totticus* A. Weiss; *29* – *Entosphaeroides* sp.; *30* – *Stiganema*-like forms; *31* – *Eosynechococcus* spp.; *32* – *Archaeotrichion contortum* Schopf; *33* – *Eomycetopsis robusta* Schopf emend. Knoll, Golubic; *34* – *Leiotrichoides typicus* Herm. and/or *L. gracilis* Pjat.; *35* – *Asperatofilum experatus* (Herm.); *36* – *Taenitrichoides jaryschevicus* Assejeva; *37* – *Obruchevella cylindrica* (Tynni et Donner) emend. Jank.; *38* – *Siphonophycus atenuatum* A. Weiss; *39* – *S.* sp.; *40* – *Rugosoopsis tenuis* Tim. et Herm.; *41* – *Tortunema sibirica* Herm. emend. Herm., *Glomovertella eniseica* (Herm.) and/or *G. glomerata* (Jank.); *42* – *Rectia costata* (Jank.); *43* – *Polytrichoides lineatus* Herm.; *44* – *Archaeoellipsoides* (=*Brevitrichoides*) *grandis* Hor. et Donald.; *45* – *Plicatidium latum* Jank.; *46* – *Cephalophytarion variabilia* Schopf; *47* – *Botuobia magna* (Tynni et Donner) emend. Kolosov; *48* – *B. patomiva* Kolosov; *49* – *Palaeolyngbya catenata* Herm. and/or *P.* sp.; *50* – *Oscillatoriopsis media* Mend. et Schopf; *51* – *Calyptothrix* sp.; *52* – *Trachytrichoides ovalis* Herm.; *53* – *Arctacellularia kelleri* Herm. et Jank.; *54* – *A. ?doliiformis* Herm.; *55* – *Oscillatoria*-like trichomes, including those with intercellular inclusions; *56* – large trichome-like structures (width 140 - 150 μm and more); *57* – *Caudosphaera expansa* Herm. et Timof.; *58* – *Germinosphaera unispinosa* Mikh.; *59* – *Ulophyton ripheicum* Tim. et Herm. and/or *Majaphyton antiquam* Tim. et Herm.; *60* – cuticular tubular coarse-costate structures; *61* – ribbonlike structureless thalluses, up to 50-μm wide, *62* – the same, up to 100-μm wide; *63* – the same, up to 200-μm wide; *64* – ribbonlike filaments with dark longitudinal zones, up to 10-μm wide; *65* – the same, up to 25-μm wide; *66* – the same, up to 100-μm wide; *67* – spindle-like chamber structure vesicles ex gr. *Brevitrichoides tiulmenicus* Jank. et Kozlova; *68* – fragments or broken remnants of forms of specific structure.

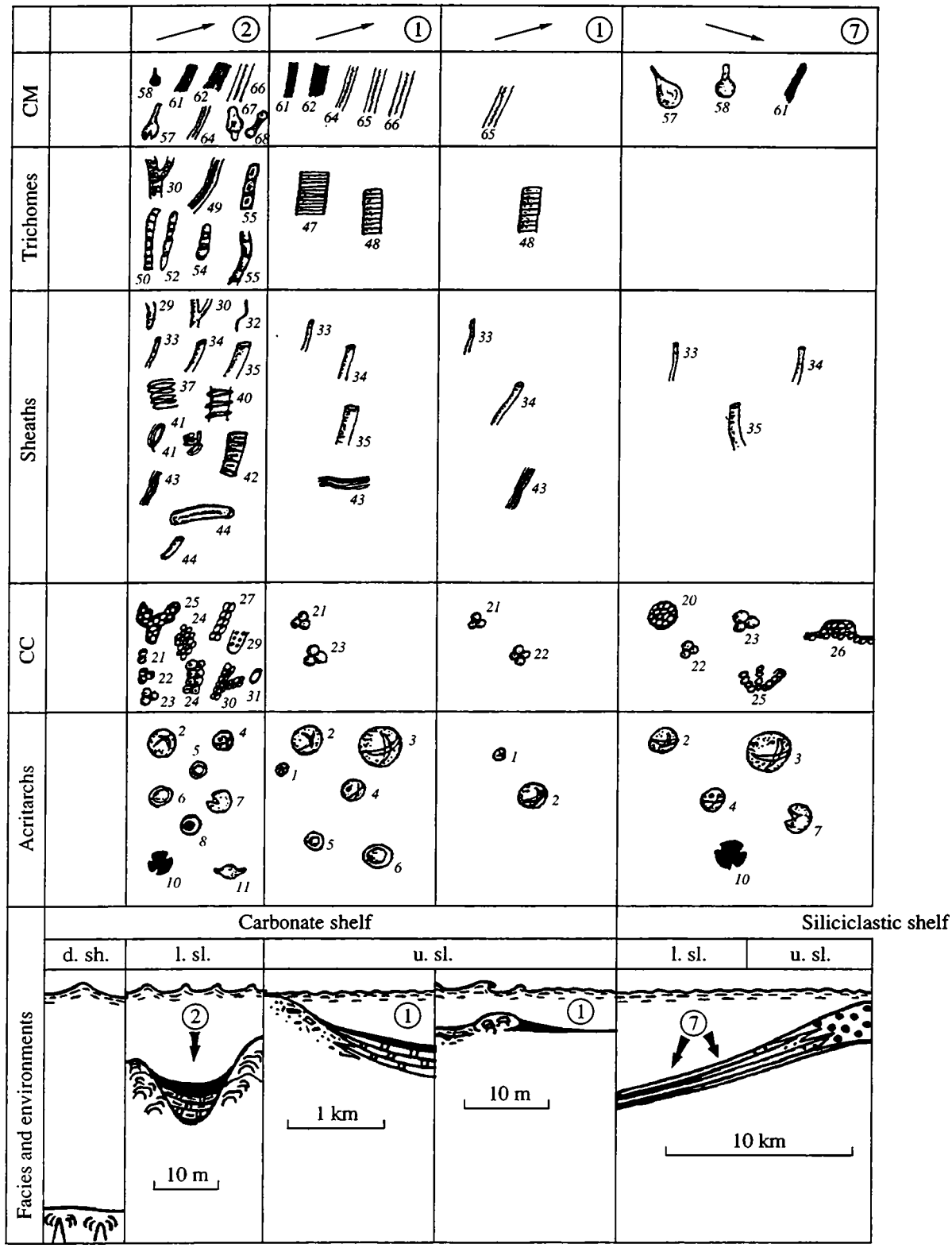
CM
Trichomes
Sheaths
CC
Acritarchs
Facies and environments
Carbonate shelf
Siliciclastic shelf
d. sh.
l. sl.
u. sl.
l. sl.
u. sl.
10 m
1 km
10 m
10 km

Fig. 5.

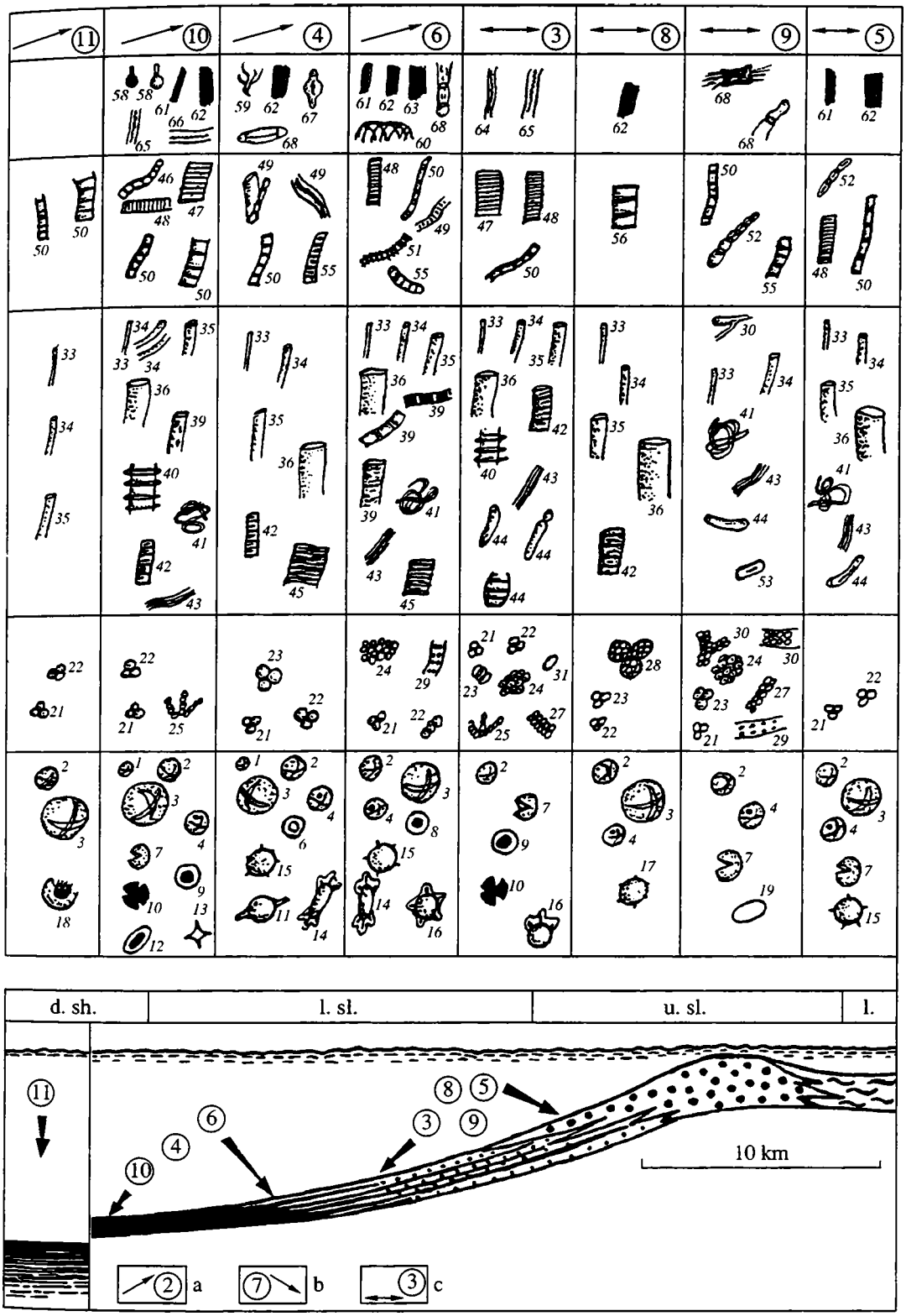

Fig. 5. (Contd.)

that are energetically favorable for phototrophic organisms. However, not being absolute, this barrier admitted the invasion of the complex morphotypes or their assemblages into the shallow parts of the shelf. However, the realization of this possibility required the appearance of some local zones of sharply reduced hydrodynamic activity within the "forbidden" areas. These local zones correspond most frequently to some floor areas with specific relief. The Derevnya shelf had three categories of these areas. The smallest of them (from a few meters to several dozen of meters across) corresponded to the interbiohermal depression traps (level 2) or to the areas in the wave shade of stromatolite bioherms (the upper sample from level 1). The larger sites (from several hundreds of meters to a few kilometers across) represented the semiisolated lagoons behind the biohermal buildups and/or carbonate-sandy shoals (the lower sample from level 1). The most extensive areas, which probably reached several kilometers across, were the quiet-water parts of the inner protected shelf (levels 3, 5 and 7 - 9).

These speculations may explain the untypical (relatively shallow-water) location of the "optimal" grouping in the Derevnya shelf. Thus, all discussed biotas, except that of the seventh level, can be considered as belonging to the same "optimal" grouping and some of them (present at levels 4, 6, and 10) occupied the typical bathymetric position and were confined to the lower subtidal zone. In contrast, the assemblages of levels 1, 3, 5, 8, and 9 occupied the shallow parts of the Derevnya shelf, i.e., the entire area of upper subtidal zones, beginning from its boundary with the lower subtidal zone. However, their penetration into this area required the occurrence of the above-mentioned specific shallow channels with low hydrodynamic activity.

On the contrary, the taxonomically reduced assemblage of the eleventh level originated in the deep shelf zone, but its impoverishment might have been caused by taphonomic factors unrelated to its original composition. As for the assemblages of the seventh level, it can be referred, with some reservation, to the transitional facial-ecological grouping of the Riphean microorganisms.

The discussed bathymetric distribution pattern of the microfossil assemblages is emphasized by the proportions of definite morphotypes in their composition. By habitation depth, all of the found forms can be conventionally divided into two groups: very few "eurybionthic" and abundant "stenobionthic" forms. The first group consists only of small *Leiosphaeridia* (diameter up to 70 - 100 μm); fragmental *Myxococcoides* (the cell size up to 15 - 20 μm); and isolated sheaths of *Eomycetopsis* and *Leiotrichoides* (2 - 20 μm wide), which are known from the entire range of the Derevnya shelf paleoenvironments. In the form of extensive interlacing mats (prominent representatives of the benthos), these sheaths were confined to more restricted facial conditions, namely, to deeper and/or more quiet environments. Most of the characteristic planktonic and benthic elements of the Derevnya microbiota show a similar selective distribution; thus, they belong to the "stenobionthic" group. The predominant part of this group (more than 90% of genera and species), especially all of the largest and complex forms, were confined to the medium-depth interval corresponding to the lower subtidal zone (Fig. 6). Many of these forms probably penetrated into the deeper-water areas of the Derevnya shelf as well, but the available material neither supports nor invalidates this assumption.

The set of taxa inhabiting the shallowest settings of the upper subtidal zone can be defined with confidence. These are the following: almost all Derevnya acritarchs, including *Chuaria* (up to 600 - 800 μm), *Trachyhystrichosphaera*, *Pterospermopsimorpha*, *Konderia*, *Simia*, and *Nucellosphaeridium*; *Ostiana*; *Eosynechococcus*; smooth-walled and wrinkled sheaths (90 - 120 μm wide); *Rectia*; *Polytrichoides*; *Glomovertella*; branching *Stiganema*-like remnants; trichomes of *Oscillatoriopsis*; broad *Botuobia* (up to 140 - 150 μm); large *Archaeoellipsoides* (= *Brevitrichoides*); fungilike *Germinosphaera*; thalluses (up to 100 μm wide), including those with dark longitudinal zones; and some other forms (Fig. 6). The set includes representatives of undoubtedly planktonic and benthic forms. As noted above, the penetration of both of these forms into the hydrodynamically active upper subtidal zone can be explained by the local appearance of quiet-water environments typical of all forms listed above.

Most of the representative Derevnya morphotypes are confined to the transgressive successions (levels 1, 2, 4 - 6, 10, and 11). In our material, these are the largest *Chuaria* and the broadest smooth-walled and wrinkled filamentous sheaths and thalluses, as well as *Obruchevella*, *Plicatidium*, *Brevitrichoides tjulmenicus* Jank. et Kozl., *Arctacellularia*, *Trachytrichoides*, *Ulophyton*, *Majaphyton*, *Archaeoclada*, and fungilike structures. The list of forms found in the regressive successions is limited to *Oscillatoriopsis*, *Eosynechococcus*, fragments of large trichomelike structures, and palmelloidal colonies. The microfossils occurring mainly in the sediments of the stable periods of the basin development comprise coccoidal *Conjunctiophycus*, *Ostiana*, *Palaeopleurocapsa* and the medium-sized thalluses and sheaths. It is meaningfull, that confinement to particular developmental trends of the basin is equally typical of both planktonic and benthic forms. It can be speculated that the contribution of biologically active compounds to the paleobasin from the provenience during the early phases of transgressions might be responsible for the assemblage abundance and diversity. Similar ideas on the influence of nutrient supply on Precambrian biota diversity have already been advanced in the literature (Vidal, 1976; Vidal and Knoll, 1983).

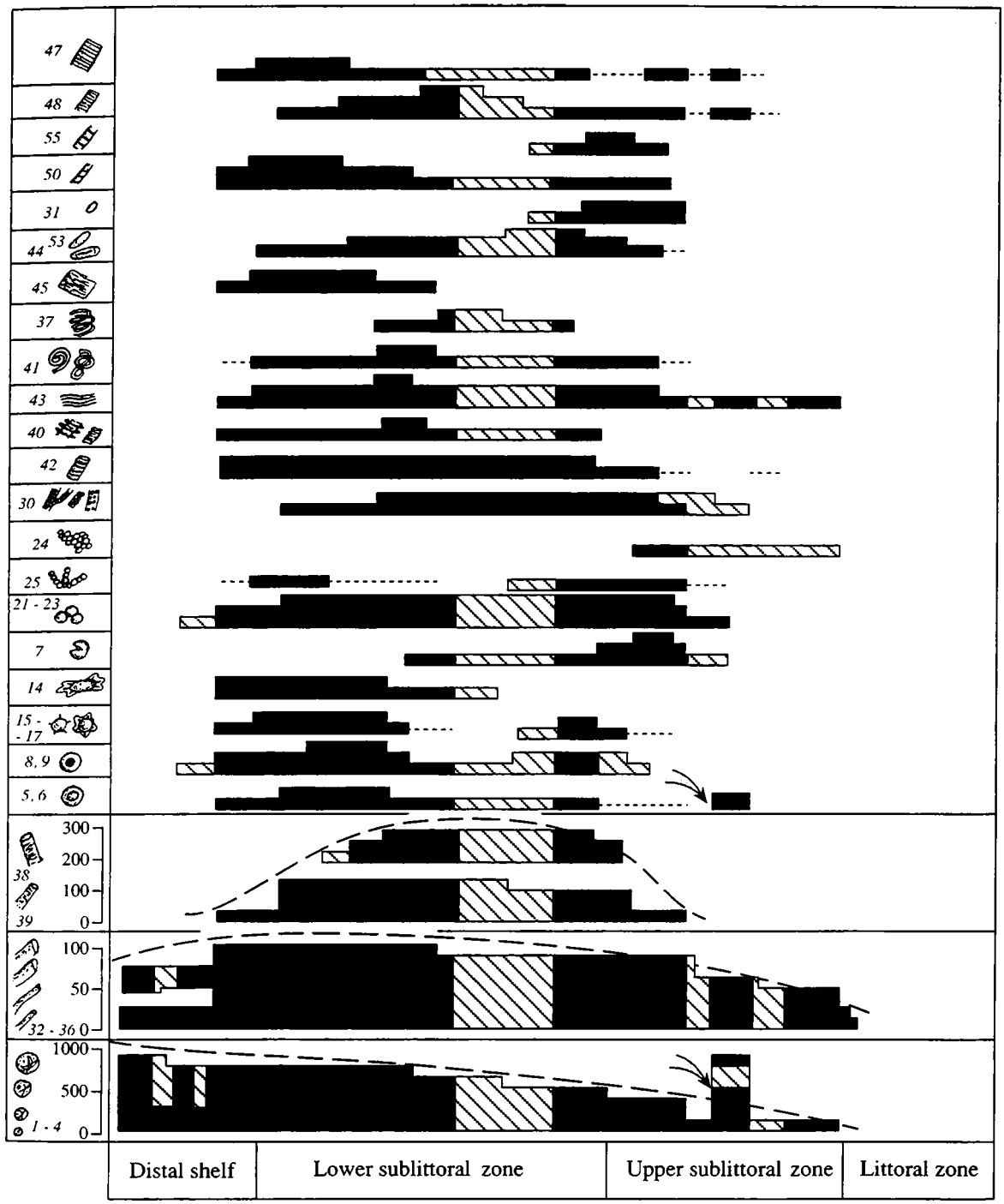

Fig. 6. Occurrence frequency (upper diagrams) and dimensional characteristics (lower diagrams vertical scale in μm) of main microfossil morphotypes and their relation to the Derevnya shelf paleoenvironments.

Stripe width indicates specimen abundance in three conventional gradations: single occurrences, common component, abundant microfossils. Black color – observable; cross hatching – inferred. Arrows show the influence of the allochthonous component (plankton) on the general sample composition. Numbers at morphotype drawings denote the microfossil taxa (see Fig. 5).

The changes in composition between the Derevnya microorganisms from the open outer and inner protected shelf appear to be informative as well. The diversity and sizes of the Precambrian microfossils (more exactly, planktonic forms) are commonly considered to increase in the offshore direction, toward the open-sea paleoenvironments (Knoll, 1982, 1984; Strother *et al.*, 1983; Knoll *et al.*, 1991; and others). However, the Derevnya microfossils, which correspond, according to Knoll's model, to relatively deep-water forms, display

a more complicated distribution pattern in the basin. Among them, there are taxa confined either to open or to protected shelf areas. The open-shelf inhabitants include *Obruchevella*, broad wrinkled sheaths (200 μm and more across), the largest *Botuobia* (up to 140 - 150 μm wide), and swell-body thalluses similar to *Archaeoclada*; all of them are benthic forms. Inhabitants of the protected shelf comprise all acanthomorphic acritarchs *Trachyhystrichosphaera* and *Prolatoforma*, entophysalean and pleurocapsian colonies of the coccoidal forms, broad (up to 140 μm) smooth-walled sheaths, filamentous *Plicatidium*, ellipsoidal *Eosynechococcus*, large trichomelike structures (up to 140 μm wide), branching thalluses of *Ulophyton*, the broadest thalluses with longitudinal structure, and some cuticular remnants with wrinkled-costate structures. Among the listed forms, there are, undoubtedly, planktonic and benthic representatives. Thus, the microfossil distribution pattern in the Derevnya shelf is consistent with Knoll's model to a certain extent, though it contradicts it in some details, e.g., acritarchs of the lower subtidal zone of the open shelf (the second level) were less diverse here than in the upper subtidal zone (the first level).

To evaluate our statements, it should be taken into consideration that specific features of the Derevnya basin predetermine the limited and specific character of the available material. Its most noteworthy features were the following: (1) the confinement of the open-sea Derevnya assemblages to somewhat atypical settings, such as some local protected sites in the shallow-water zones; (2) the impoverished microfossil assemblages of the open basin parts, which left incomplete and, to some degree, occasional characteristics of these areas; (3) the absence of organic-walled microfossils in the deepest open environments represented by the rhythmical biostrome facies; and (4) the abundance of stromatolites in these environments that *a priori* increases a portion of the autochthonous biota at the expense of other assemblages of stromatolite-forming organisms.

The appearance of large and/or complex acritarchs typical of the open-sea *optimal* medium-depth zone in the unusual semi-isolated shallow-water settings (levels 1, 5, 7, 8, and 10) could have been caused by a short-term free exchange between water masses of the open-sea and shallow proximal parts of the Derevnya basin. This exchange could only have occurred at the beginning of transgressions, when sea water almost instantly covered the wide shallow zone of the carbonate platform. The appearance of "foreign" planktonic components in the shallow areas was probably a relatively rare phenomenon resulting from the coincidence of several paleoecological factors. As the sea level and facies zones of the basin became stabilized, the taxonomic and dimensional distribution of planktonic and benthic assemblages approached the usual pattern, with decreased diversity and dimensions toward greater and lesser depths as compared with the "optimal" zone depth (levels 3, 4, 9, and 11). It is unlikely that the largest acritarchs, including their acanthomorphic forms, were able to migrate far in water column. Otherwise, it would be difficult to explain why they are so strictly confined to the size-limited "optimal" zone of the Derevnya and all basins studied earlier.

As for the benthic elements of the "optimal" grouping, they were indifferent relative to the outer and inner shelf areas, but they showed a noticeable conservatism in their colonization of the available paleoecological environments. In the studied material, the most representative benthic assemblage is confined to the second level (the open shelf, interbiohermal depression-trap). Significantly, this level yielded one of the poorest acritarch assemblages. This difference in the spatial position of the diversity peaks for planktonic and benthic forms can be regarded as an argument in favor of the original character of the lateral structure of the Precambrian biotas described earlier (Veis and Petrov, 1994a, 1994b). The other assemblages of benthic forms of the Derevnya Formation originated in the deep distal part of the inner shelf (the sixth and tenth levels) or in the near-shore quiet zone between the lower and upper subtidal zones (the ninth level). They are almost as diverse as the assemblage of the second level. In general, all benthic forms of the Derevnya biota tend to be concentrated in quiet, relatively deep-water shelf paleoenvironments.

Thus, using one of the richest Late Riphean microbiotas as an example, we confirmed our previous conclusion that the depth factor was of prime importance for the lateral distribution of particular microfossil assemblages along the basin profile. We also revealed that the distribution was influenced by some other factors, the most significant of which, were hydrodynamic activity and the confinement of particular biota to certain stages (transgressive, regressive, or stable) of the general evolutionary trend of the basin. In other words, the distribution of taxa and assemblages of the Riphean microfossils in time and space was dependent on a number of parameters and was more complicated than is commonly appreciated in the current literature.

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Carbon and Oxygen Isotopic Variations in the Lower Proterozoic Carbonates of the Kola Peninsula

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Abstract – Significant time-dependent variations of $\delta^{13}C$ were distinguished in the Lower Proterozoic carbonates of the Pechenga–Varzuga belt. The Jatulian rocks (2.3 - 2.1 Ga) revealed unusually high values of $\delta^{13}C$ (up to 7.4), which suggests that they evidently originated in shallow continental protoevaporite basins. The post-Jatulian rocks showed highly variable values of both $\delta^{13}C$ (–12.5 - 2.9‰) and $\delta^{18}O$ (9.6 - 21.4‰). This is indicative of an intense postsedimentary modification of the carbonate composition under the influence of fluids rich in CO_2 of organic origin. The changes in isotopic composition were associated with the carbonate enrichment in strontium (up to 1900 ppm in samples with higher concentrations of light isotopes) and, to some extent, in manganese.
Key words: **Lower Proterozoic, carbonates, carbon and oxygen isotope composition.**

In general, the Archean and Proterozoic sedimentary carbonates are similar in their C-isotope composition to recent analogues – $\delta^{13}C = 0 \pm 2$ - 4‰ (Keith and Weber, 1964; Shidlowski *et al.*, 1975, 1983). Because this parameter is under the strong control of carbon distribution between the organic and nonorganic carbonate reservoirs, its constancy is believed to be indicative of the high stability of carbon cycles and the very early origin of life on Earth in a volume comparable to the present one.

However, Phanerozoic and Precambrian deposits are known in which carbon has an anomalous isotopic composition. These anomalies may be related to either abnormal local conditions of sedimentation or global changes in the carbon cycles.

An abnormally heavy isotopic composition of carbon has been measured in the following Lower Proterozoic rocks: the 2.65- to 1.95-Ga-old carbonates of the Lomagundi province in Rhodesia, $\delta^{13}C = 8.2 \pm 2.6$‰ (Shidlowski *et al.*, 1975); the Jatulian (2.1 - 2.0 Ga) deposits of Finland, $\delta^{13}C = 4.0$ - 16.0‰ (Karhu, 1993); the carbonate rocks of the Karelian (2.1 - 2.0 Ga) complex, central Karelia, $\delta^{13}C = 5.6$ - 18.0‰ (Yudovich *et al.*, 1990); the Lewisian Limestone (2.0 Ga) and Lafoten carbonates (2.0 Ga) of Norway, $\delta^{13}C$ up to 12.9 and 12.1‰, respectively (Baker and Fallick, 1989a, 1989b); and the Albanel Formation (2.1 Ga) of Canada, $\delta^{13}C = 6.4$‰ (Shidlowski *et al.*, 1983).

Various hypotheses has been suggested to explain these anomalies. Shidlowski and colleagues believed that local factors were responsible for the high $\delta^{13}C$ values in carbonates of the Lomagundi province, as they had been deposited in isolated basins with high bioproductivity and stratified water masses. In the opinion of Yudovich *et al.* (1990), the positive C-isotope anomaly may indicate that the methane atmosphere of the Early Proterozoic had very low oxygen concentrations. In contrast, Tikhomirova (1991) argued that oxygen concentrations increased in the atmosphere during that time. Prilutskii *et al.* (1992) suggested that atmospheric saturation in juvenile carbon dioxide, which was rich in ^{13}C because of a high hydrocarbon concentration in fluids, was the main factor responsible for the origin of carbonates with elevated concentrations of this heavy isotope.

Analyzing the "heavy" isotopic composition in the carbonate rocks of Pechenga, Imandra–Varzuga, and the Karelian massif, Karhu and Melezhik (1992) concluded that the available data are insufficient for deciding whether their deposition resulted from local or global factors. They also noted that the synchronism of deposits with abnormally high $\delta^{13}C$ values has been inadequately proved, thus providing grounds for considering them to have been deposited under local and peculiar facial conditions. However, Karhu later (1993) returned to the idea that an intense accumulation of organic matter, global in nature, was a factor responsible for "heavy carbon" in carbonates.

In addition to rocks enriched in the heavy isotope, some Lower Proterozoic carbonates with low $\delta^{13}C$ values of an unclear origin have also been recognized among deposits of the Baltic shield.

To obtain more complete information on variations in $\delta^{13}C$ and $\delta^{18}O$ values and to analyze the factors responsible for the Early Proterozoic isotopic anomalies, we studied carbonate rocks of the Pechenga and Imandra–Varzuga structures, which form the greatest Pechenga–Varzuga greenstone belt of the Baltic shield (Fig. 1).

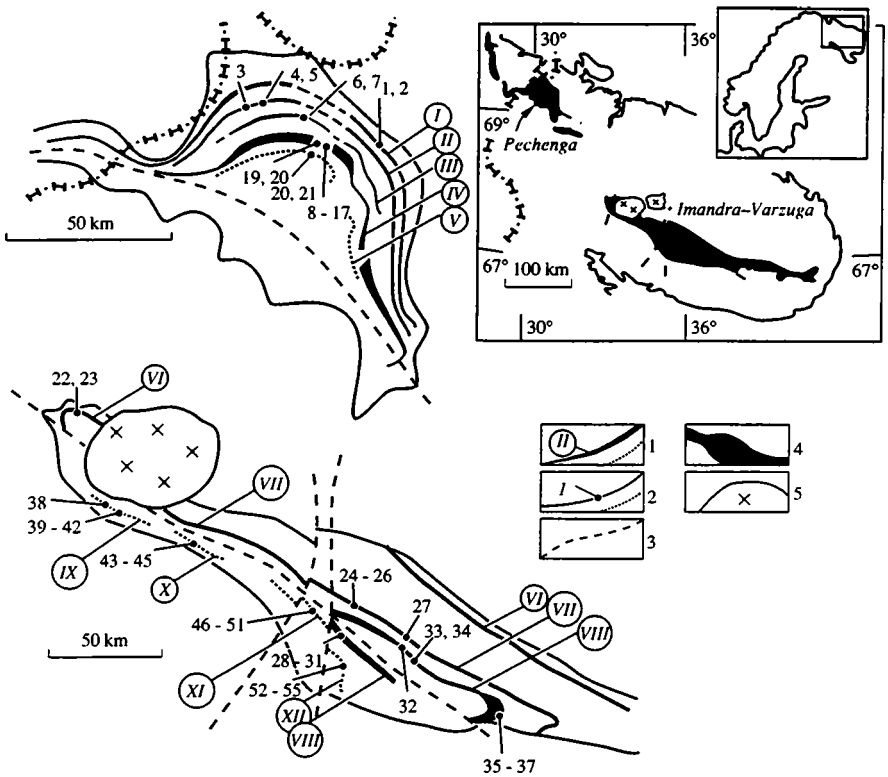

Fig. 1. Schematic maps of the Pechenga–Varzuga belt in the Kola Peninsula with indicated sampling localities.
1 – Sedimentary formations with numbers as indicated in the table; 2 – sampling localities indicated by run numbers from the table; 3 – main tectonic boundaries; 4 – positions of the Pechenga and Imandra–Varzuga zones; 5 – Khibiny alkaline massif.

MATERIALS

The stratigraphic nomenclature and correlation of volcano-sedimentary deposits in the Pechenga and Imandra–Varzuga zones that we used were adopted from published works (Zagorodnyi *et al.*, 1982; Predovskii *et al.*, 1987). The accepted geochronological scale is based on absolute age determinations obtained in the Geological Institute of the Kola Scientific Center, Russian Academy of Sciences (Fig. 2).

We studied all the levels of carbonate rocks known in the belt for the interval ranging from 2.5 - 2.4 to 2.0 - 1.9 Ga. The oldest (2.5 - 2.4 Ga) samples exemplify sandy dolomitic limestones from the Seida River Formation of the Imandra–Varzuga zone (the table and Figs. 1 and 2). These rocks usually form lenses, which are 0.2 - 0.5 m thick and occur among cross-bedding arkoses and quartz-sandstones.

The second level of carbonate rocks, which is dated at 2.3 - 2.1 Ga, includes abundant and thick dolomite beds that are traceable all along the belt. In the Pechenga zone, this level corresponds to the Kuetsyarvi Formation, the carbonate sediments of which are extremely rich in beds of dolomite, sandy dolomite, and dolomitic calcarenite. They also enclose frequent stromatolite buildups and oncolites. The thickness of these carbonate rocks ranges from to 25 to 100 m. In the Imandra–Varzuga zone, the same level is occupied by the Umba Formation with a similar rock association. However, these rocks are barren of stromatolites, but they contain numerous lenses of jasper-bearing barite. The carbonate rocks vary in thickness from 50 to 150 m.

The next level of this interval (2.3 - 2.1 Ga) is also traceable all along the belt. It is represented by thick carbonate beds of both the Kolasioki Formation of the Pechenga zone and the Il'mozero Formation of the Imandra–Varzuga zone. Here, the rocks show some differences in composition as compared to older levels (the table). Typical rocks of the Kolasioki Formation are dolomites, Mn-dolomites, and limestones associated with jasper and other siliceous beds. The thickness of the Kolasioki carbonate beds is 10 - 50 m. The Il'mozero Formation includes dominant dolomites and silicified stromatolitic dolomites. Its terrigenous rocks usually contain limy diagenetic concretions. Here, the carbonate beds are 20 - 150 m thick.

In both zones, the upper parts (2.0 - 1.9 Ga) of the sequences contain thin beds and lenses of carbonate rocks occurring among sulfide-carbonaceous shales and graywacke siltstones. This level is remarkable because of a distinct change in the general composition of the carbonate rocks, among which, limestones become dominant. Three age levels were distinguished

in the Pechenga zone in the different carbonate materials within this part of the section. The first consists of sulfide-carbonaceous graywacke siltstones bearing rare and thin (1 - 10 cm) limestone lenses. These rocks form a horizon inside the spheruloid lavas of the Kolasioki Formation. The second level is represented by the Pil'guyarvi Formation of thick (more than 1000 m) sulfide-carbonaceous deposits with numerous limestone lenses and limy diagenetic concretions in their upper beds. The horizon with these carbonate inclusions is 100 - 500 m thick, and the thickness of the lenses varies between 1 and 50 cm. The uppermost level was distinguished in the spheruloid Pil'guyarvi lavas. It consists of acid tuffs and tuffites with abundant carbonate inclusions (concretions?).

In the Imandra–Varzuga zone, the interval of 2.0 - 1.9 Ga seems to correspond to the Tominga Group of sulfide-carbonaceous shales, basic tuffs, and lavas (Zagorodnyi *et al.*, 1982; Predovskii *et al.*, 1987). Carbonate rocks form thick lenses embedded in these deposits. In the lower part of the section, the lenses of Mn-enriched "Rouksa Dolomite and Limestone" are more than 120 m thick in total, and they associate with thick beds of ferrocarbonaceous silicilith. Upward the section (localities "45 km" and Solenoe Lake), lenses of similar Mn-limestone and Mn-dolomite, occurring among highly carbonaceous sulfide shales, are thinner (0.1 - 1.0 m). Presumably, the thick (50 - 100 m) limestone lenses of the Isvestkovyi Zavod locality are the youngest in the section. They occur among basic lavas.

At all the studied levels, the carbonate rocks were metamorphosed under conditions of prehnite-pumpellyite and greenschist facies, which weakly affected the original texture and structure of the rocks.

ANALYTICAL PROCEDURE

The classical (wet) chemical method was used to determine the CaO, MgO, CO_2, and MnO contents in the studied carbonates. Sr concentrations were measured by the X-ray fluorescence method, and C_{org} was analyzed by burning the samples in a microanalytical furnace. All chemical analyses were carried out at the laboratories of the Geological Institute of the Kola Scientific Center, Russian Academy of Sciences (analysts A.N. Vakhatova for C_{org} and V.A. Pobedonostsev for Sr).

Sample decomposition for carbon and oxygen isotope analysis was performed by dissolving in H_3PO_4 according to the widely accepted technique. The isotopic composition was measured with an MI-1201B mass spectrometer. The measured values of $\delta^{13}C$ and $\delta^{18}O$ are presented as pro mille deviations relative to the standards PDB and SMOW, respectively. To relate our measurements to PDB, we used the standards KH-2 (Gerstenberger and Herrmann, 1982) and NBS-19. For recalculating $\delta^{18}O$ values relative to SMOW, we used

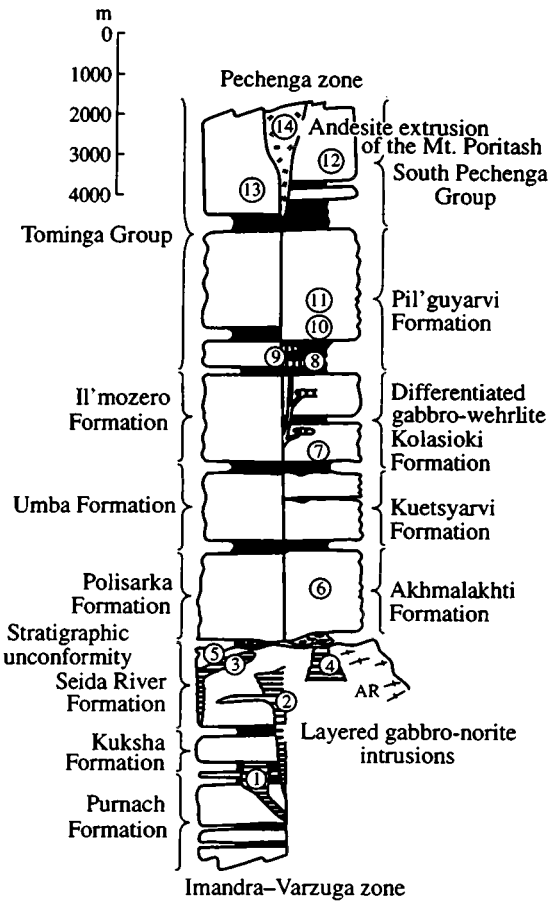

Fig. 2. Stratigraphic nomenclature, correlation, and radiometric ages of rocks in the Pechenga and Imandra–Varzuga zones.

1 – Gabbro-norite from the Monchegorsk layered intrusion: 2492 ± 31 (Sm–Nd) and 2493 ± 7 Ma (U–Pb, zircon), after Mitrofanov *et al.* (1991); 2 – gabbro-norite from layered intrusions in the Pan–Fedorov tundra: 2487 ± 51 (Sm–Nd) and 2471 ± 4 Ma (U–Pb, zircon), after Dokuchaeva *et al.* (1990) and Mitrofanov *et al.* (1991), respectively; 3 – gabbro-norite from the Imandra lopolith: 2444 ± 77 (Sm–Nd) and 2396 ± 7 Ma (U–Pb, zircon), after Balashov *et al.* (1990) and Mitrofanov *et al.* (1991), respectively; 4 – gabbro-norite from the layered intrusion of the Mt. General'skaya: 2453 ± 42 Ma (Sm–Nd), after Bakushkin *et al.* (1990); 5 – acid volcanic rocks of the Seida River Formation: 2423 ± 7 Ma (U–Pb, zircon), after Mitrofanov *et al.* (1991); 6 – basaltic andesite of the Akhmalakhti Formation: 2324 ± 28 Ma (Rb–Sr), after Balashov *et al.* (1990); 7 – tholeiite basalt of the Kolasioki Formation: 2114 ± 52 Ma (Rb–Sr) after Balashov *et al.* (1991); 8 – gabbro-wehrlite intrusions: 1990 ± 55 (Pb–Pb), 1980 ± 72 (Rb–Sr), and 1970 ± 110 Ma (Sm–Nd), after Mitrofanov *et al.* (1990); 9 – sulfide Cu–Ni-ore from gabbro-wehrlite: 1870 ± 50 Ma (Pb–Pb), after Mitrofanov *et al.* (1990); 10 – ferropicrite of the Pil'guyarvi Formation: 1955 ± 43 (Pb–Pb) and 1990 ± 66 Ma (Sm–Nd), after Hanski *et al.* (1990); 11 – rhyolite tuff of the Pil'guyarvi Formation: 1970 ± 5 Ma (U–Pb, zircon), after Hanski *et al.* (1990); 12 – andesite from the South Pechenga complex: 1855 ± 54 Ma (Rb–Sr), after Balashov *et al.* (1991); 13 – acid volcanic rocks of the Tominga Formation: 1870 ± 71 Ma (Rb–Sr), after Mitrofanov *et al.* (1991); 14 – andesite from the Poritash subvolcanic intrusion: 1729 ± 35 Ma (Rb–Sr), after Balashov *et al.* (1991).

Carbon and oxygen isotopic parameters and chemical composition of carbonates from the Pechenga–Varzuga complex

Run	Sample	Rock type	^{13}C, ‰ (PDB)	^{18}O, ‰ (SMOW)	CO_2, mass %	MgO, mass %	CaO, mass %	MnO, mass %	C_{org}*, mass %	Sr, ppm	C_{org}**, mass %
				Pechenga complex							
				I. Kuetsyarvi Formation							
1	P-000A	Sandy dolomite	+6.8	+18.8	31.39	14.54	20.74	0.01	0.02	n. d.	0.05
2	P-000B	"	+7.4	+18.2	33.34	15.17	22.62	0.02	0.01	n. d.	0.05
				II, III. Kolasioki Formation							
3	C-18/108	Limestone	−5.3	+12.8	36.88	4.56	41.05	0.56	0.03	153	0.55
4	C-29/410	Sandy limestone	−5.5	+13.1	25.69	2.70	29.66	0.52	0.02	183	0.55
5	C-29/411.5	Dolomite	−5.0	+14.1	33.04	13.72	25.12	1.08	0.01	120	0.55
6	28/54	Sandy limestone	−7.8	+11.5	23.27	1.21	30.58	0.11	0.03	102	0.8
7	28/69	Limestone	−7.6	+10.2	35.82	0.38	46.96	0.19	0.22	188	0.8
				IV, V. Pil'guyarvi Formation							
8	K-2900/561.8	Late-diagenetic calcareous concretion	−7.5	+11.6	30.15	1.86	38.97	0.16	0.42	296	0.6
9	K-2900/563.9	"	−8.2	+13.0	32.10	1.83	41.08	0.12	0.35	325	0.6
10	K-2900/580.4	"	−10.3	+12.1	29.83	2.71	39.22	0.33	0.52	767	0.6
11	2900/721	"	−13.3	+12.3	15.78	2.62	21.45	0.14	0.18	523	0.6
12	2900/1009	Carbonaceous limestone	−8.7	+12.7	30.86	0.81	39.18	0.18	1.39	1689	2.0
13	2900/1011	"	−8.7	+12.4	32.85	0.78	40.56	0.23	1.04	1846	2.0
14	2900/1075	"	−12.5	+12.7	31.06	0.69	39.76	0.37	0.40	1865	2.0
15	2900/535oz	Outer zone of late-diagenetic carbonate concretion (LCC)	−13.2	+12.3	5.66	0.67	7.78	0.04	n. d.	218	0.6
16	2900/535mz	Middle zone of LCC	−14.1	+10.8	20.96	0.84	26.74	0.12	n. d.	779	0.6
17	2900/535c	Core of LCC	−13.1	+12.3	30.24	1.07	38.35	0.14	n. d.	1074	0.6
18	2400/248	Late-diagenetic carbonate concretion	−11.3	+11.2	26.19	0.66	32.83	0.34	0.25	984	0.6
19	2798/600	Spherolitic carbonaceous limestone	−9.7	+12.6	30.96	0.75	40.02	0.16	0.65	1415	2.0
20	SGS-3oz	Outer zone of diagenetic carbonate concretion (DCC)	−7.1	+11.8	30.58	0.66	44.05	0.56	0.06	116	0.03
21	SGS-3c	Core of DCC	−7.9	+9.6	31.61	0.58	45.83	0.50	0.07	128	0.03
				Imandra–Varzuga complex							
				VI. Seida River Formation							
22	4757		−4.8	+13.9	18.41	2.49	19.09	0.08	0.01	487	0.15
23	4758b		−4.0	+14.5	29.50	3.72	33.81	0.12	0.01	630	0.10
				VII. Umba Formation							
24	381	Dolomite	+2.3	+13.8	40.68	15.67	31.63	0.12	0.01	361	0.05
25	823	"	+6.4	+13.8	37.98	14.27	29.96	0.10	0.02	283	0.02
26	182	"	+4.5	+11.0	39.33	15.11	29.55	0.10	0.01	257	0.04
27	486	"	+4.0	+15.7	36.93	14.71	29.81	0.11	0.01	180	0.03

(Contd.)

Run	Sample	Rock type	^{13}C, ‰ (PDB)	^{18}O, ‰ (SMOW)	CO_2, mass %	MgO, mass %	CaO, mass %	MnO, mass %	C_{org}*, mass %	Sr, ppm	C_{org}**, mass %
			VIII. Il'mozero Formation								
28	154/659	Sandy dolomite	−0.5	+21.4	37.0	16.06	25.21	0.03	n. d.	62	0.3
29	154/660	"	−0.8	+21.3	25.45	11.50	17.50	0.03	n. d.	82	0.3
30	154/662	Limestone	−0.3	+19.2	37.15	4.49	42.23	0.09	n. d.	134	0.3
31	154/663	"	−0.4	+19.0	38.60	6.76	40.46	0.09	n. d.	101	0.3
32	415	"	+2.0	+17.7	39.66	0.66	50.60	0.11	0.01	210	0.04
33	469-k	Diagenetic carbonate concretion	−3.9	+13.4	20.20	1.00	27.93	0.61	n. d.	406	0.04
34	877-k	Diagenetic calcareous concretion	−4.1	+12.8	16.02	1.38	22.43	0.47	n. d.	350	0.04
35	735	Limestone	+2.0	+18.6	37.06	3.36	47.12	0.11	0.01	209	0.04
36	736	"	+2.0	+18.6	35.62	5.52	44.04	0.12	0.01	190	0.04
37	3027-k	Diagenetic calcareous concretion	−20.5	+13.3	21.76	2.99	28.76	0.36	n. d.	463	0.4
			IX - XII. Tominga Formation								
38	NIZ-1		+2.9	+24.1	n. d.	n. d.	n. d.	n. d.	0.01		0.02
39	540	Limestone	+0.7	+11.5	37.97	0.91	49.05	0.08	0.01	136	0.02
40	544	"	+1.1	+19.2	36.83	0.87	48.75	0.10	0.01	120	0.02
41	545	"	+2.4	+14.9	34.54	0.40	45.37	0.06	0.01	132	0.02
42	547	"	+1.0	+13.7	36.12	0.95	45.82	0.10	0.01	471	0.02
43	890	Limy dolomite	−5.0	+15.1	34.74	12.32	25.37	0.12	0.20	293	1.0
44	909	"	−2.6	+16.6	31.01	10.68	21.15	0.15	0.42	306	1.0
45	935	"	−2.7	+16.9	30.66	11.33	20.84	0.39	0.67	235	1.0
46	87	Sandy dolomite	−9.7	+16.1	20.43	11.15	13.40	0.37	0.28	315	2.0
47	100	Sandy calcareous dolomite	−7.5	+12.2	30.25	10.87	20.96	0.46	1.11	339	3.0
48	110-A	Sandy limestone	−12.5	+15.1	19.77	2.71	22.06	0.31	0.86	609	3.0
49	112	Sandy calcareous dolomite	−7.3	+16.6	29.71	10.14	23.68	0.60	0.16	403	3.0
50	153	"	−7.0	+16.9	33.48	11.52	22.53	0.38	1.00	293	3.5
51	157	Sandy dolomite	−11.8	+16.2	29.67	11.23	19.22	0.40	0.91	493	3.5
52	570	"	−0.7 (−0.1)	+18.2 (+17.4)	38.89	15.87	26.12	0.10	0.10	262	0.09
53	571	Dolomite	+0.1 (0.8)	+19.5 (19.7)	40.16	16.35	28.31	0.09	0.09	275	0.08
54	576	Calcareous dolomite	−1.4	+19.8	36.96	11.91	24.74	2.16	0.46	263	0.4
55	590	"	−0.4 (0.1)	+18.1 (18.9)	42.37	14.94	33.98	0.12	0.09	327	0.09

Notes: * C_{org} in carbonates.

** C_{org} in rocks enclosing carbonate layers and concretions.

Numbers of runs and formations are those shown in Fig. 1.

Values in parentheses for dolomite, others for calcite.

the following equation: $\delta^{18}O_{SMOW} = 1.03086\delta^{18}O_{PDB} + 30.86$ (Friedman and O'Neil, 1977). The analytical uncertainty is below ±0.2% for both $\delta^{13}C$ and $\delta^{18}O$.

RESULTS

Carbonate rocks in both the Pechenga and the Imandra–Varzuga zones show highly variable values of $\delta^{13}C$ and $\delta^{18}O$. In the Pechenga zone, $\delta^{13}C$ ranges from −14.1 to 7.8‰, and $\delta^{18}O$ values are between 9.6 and 20.3‰. In the Imandra–Varzuga zone, these parameters vary respectively from −20.5 to 6.7‰ and from 11.5 to 23.1‰ (the table).

Analyzing the variations in carbon isotope composition throughout the sections, we noted the following regularities: (1) the pre-Jatulian (Seida River) formation of dolomitic limestones (2.5 - 2.4 Ga) has moderately low $\delta^{13}C$ (from −4.8 to −4.0‰), which agrees well with the data of Karhu and Melezhik (1992); (2) the lower Jatulian carbonates (mostly dolomites) in both the Pechenga (Kuetsyarvi Formation) and the Imandra–Varzuga (Umba Formation) zones show abnormally high $\delta^{13}C$ (2.3 - 7.4‰); (3) the upper and post-Jatulian (Lyudy Superhorizon) complex of limestones and some dolomites (2.1 - 1.9 Ga) reveals abnormally low $\delta^{13}C$ (from −5.0 to −12.5‰) in the Pechenga zone, while in the Imandra–Varzuga zone, the respective parameters are either normal (approaching zero) or low, being extremely variable (from −12.5 to 2.9‰) in rocks of the Tominga Group; and (4) diagenetic calcareous concretions from different age levels have low and highly variable $\delta^{13}C$ (ranging between −3.9 and −20.5‰), which is typical of such carbonate segregations.

We distinguished a negative correlation between the $\delta^{13}C$ values in carbonates and the C_{org} concentrations in the rocks enclosing carbonate interlayers (Fig. 4). This interdependence become less evident when we deal with both of the parameters measured in the carbonates proper. All abnormally low values of $\delta^{13}C$ (except for two samples from the Tominga Group) were measured in horizons, where concentrations of C_{org} are above 0.5%. Similarly, concretions from carbon-free beds ($\delta^{13}C$ ranges from −3.9 to −7.9‰) differ from those enclosed in carbonaceous rocks ($\delta^{13}C$ ranges from −7.5 to −20.5‰). In a single bed series, which comprises different sedimentary facies, concretions become enriched in light carbon isotope when they occur in coastal carbon-free deposits, but not in carbonaceous sediments of deeper origin (Fig. 5).

In limestones with abnormally low values of $\delta^{13}C$, we measured unusually high (up to 1865 ppm) Sr concentrations (the table). In some sections, for instance, the Tominga Group, there is a negative correlation between $\delta^{13}C$ values and Sr concentrations (Fig. 6). However, deposits with a normal range of $\delta^{13}C$ values, e.g., those of the Il'mozero Formation, show an opposite trend of balanced changes in Sr concentrations and $\delta^{13}C$ values.

There is a positive correlation between $\delta^{13}C$ and $\delta^{18}O$ values (Fig. 7). Rocks with normal and high values of $\delta^{13}C$ have an oxygen isotope composition that is typical of Precambrian carbonates. However, limestones with abnormally low values of $\delta^{13}C$ are distinctly depleted in ^{18}O ($\delta^{18}O$ up to 9.6‰), which suggests that they suffered a considerable loss of original O-isotope characteristics.

DISCUSSION

The carbon isotope composition in carbonate rocks is dependent on three main factors: (1) the isotopic exchange between the CO_2 dissolved in the water and the CO_2 of the atmosphere, (2) the consumption of CO_2 enriched in light carbon isotope by organisms for photosynthetic reactions, and (3) the reflux of CO_2 into the water after the oxidation of organic matter rich in light carbon isotope. It appears that the isotopic exchange between water and the atmosphere is hardly dependent on facial environments. This is the factor that usually levels the isotopic composition of carbonates, regardless of their depositional setting. In contrast, the rates of photosynthesis and oxidation of organic materials are highly dependent on local environments. Their changes are responsible for a wide spectrum of isotopic anomalies observable in recent and ancient carbonate sediments. Therefore, the results of paleofacial analysis are very significant for a correct interpretation of isotopic data.

So far, our results for the pre-Jatulian rocks (Seida River Formation) are limited and not very useful for a detailed discussion. We note only that the pre-Jatulian dolomitic limestones (2.5 - 2.4 Ga) were deposited in fresh-water lake basins, as follows from the available lithologic and geochemical data (Melezhik, 1987), and that this was the most suitable factor that could have been responsible for the low $\delta^{13}C$ values in these rocks. For instance, carbonates accumulating in many recent flow-through lakes have a similar isotopic composition (Talbot, 1990). However, it is not inconceivable that the isotopic composition in the pre-Jatulian carbonates could have been distorted by diagenetic and catagenetic processes.

The Jatulian dolomites and associated limestones (2.3 - 2.1 Ga) revealed a wide range of predominantly high $\delta^{13}C$ values (0 - 7.4‰). Taking into account the data on adjacent regions (Yudovich *et al.*, 1990; Karhu and Melezhik, 1992; Karhu, 1993), where carbonates with high $\delta^{13}C$ also dominate at approximately the same level, we may consider the Jatulian anomaly to be typical of the entire Baltic shield (Fig. 8).

The reasons for the anomalous enrichment of carbonates in ^{13}C are still poorly understood. It is likely that two factors are important here: (1) the high photosynthesis ability of organisms and (2) the stratification of water masses, preventing the oxidation of organic substances and the resaturation of the surface water

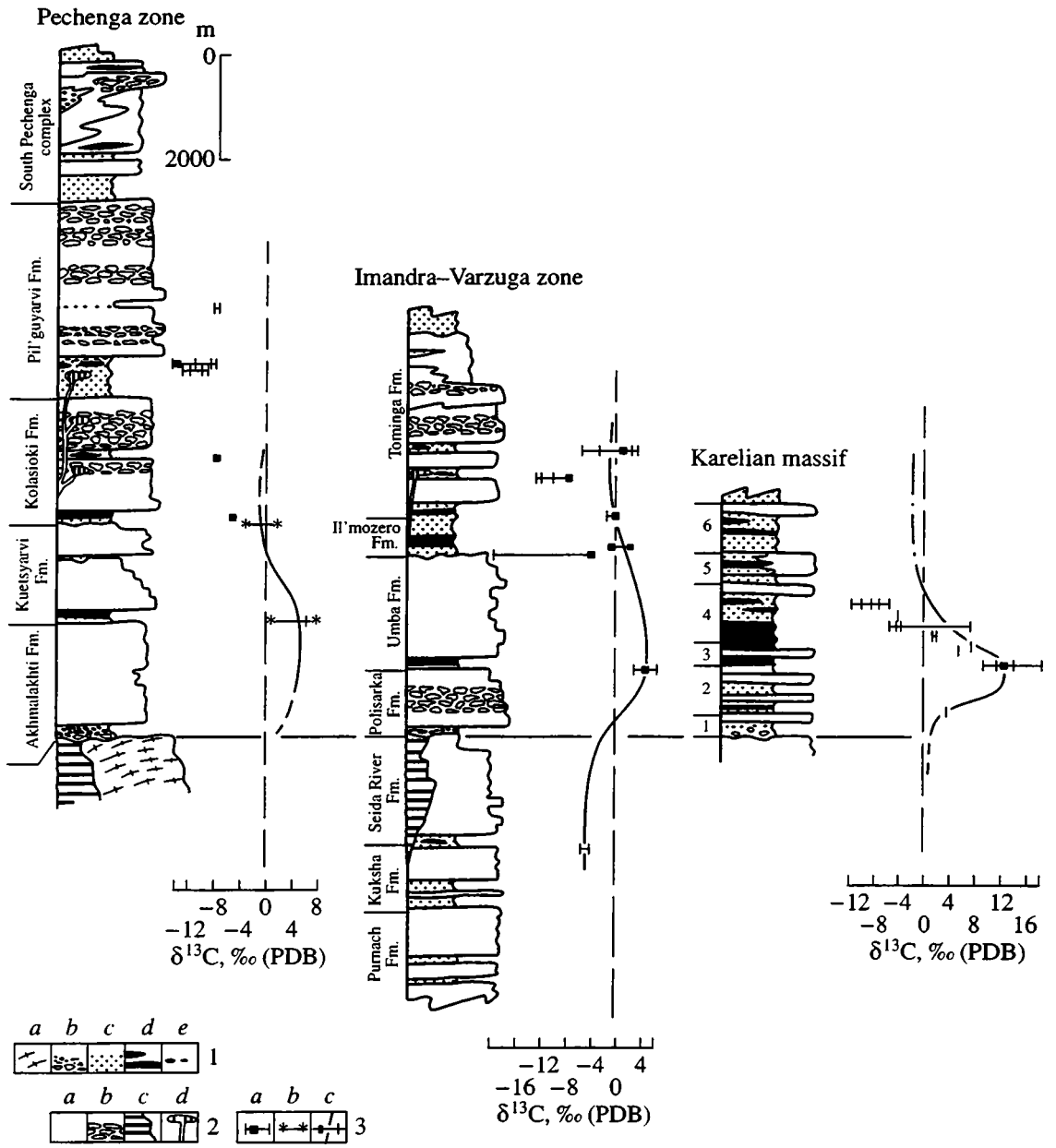

Fig. 3. Variations of carbon isotope composition in carbonates of the volcano-sedimentary sequences.
1 – Archean granite-gneisses (*a*), Lower Proterozoic conglomerate (*b*), other sedimentary rocks (*c*), carbonates (*d*), carbonate concretions (*e*); 2 – Lower Proterozoic volcanic rocks (*a*), levels of pillow-lavas (*b*), layered gabbro-norite intrusions (*c*), differentiated Ni-bearing intrusions of the Pechenga type (*d*); 3 – data after Karhu and Melezhik (1992) and from this study (*a*), data after Prilutskii *et al.* (1992) (*b*), connecting lines for median values (*c*). Data for the Karelian massif are cited from Yudovich *et al.* (1990); numbers indicate the following horizons: Sarioly (1), Segozero (2), Onega (3), Lyudy (4, 5), and Kalevy (6). Stratigraphic subdivisions for the Pechenga and Imandra–Varzuga zones are after Zagorodnyi *et al.* (1982) and Predovskii *et al.* (1987), and those for the Karelian massif are after Early (1986).

layer with carbon dioxide rich in the light C-isotope. Both conditions are characteristic of closed saline and hypersaline basins.

The high enrichment in ^{13}C discovered by Shidlowski *et al.* (1984) in microbial assemblages inhabiting coastal brines on the Sinai Peninsula are very remarkable. This seems to be a result of low CO_2 solubility in highly concentrated brines, where photosynthesis is also peculiar. The real significance of both factors in causing carbonate enrichment in heavy C-isotope is insufficiently understood; however, it is remarkable that the Jatulian rocks from the Karelian zones also contain organic substances with high $\delta^{13}C$ (Yudovich *et al.*, 1990).

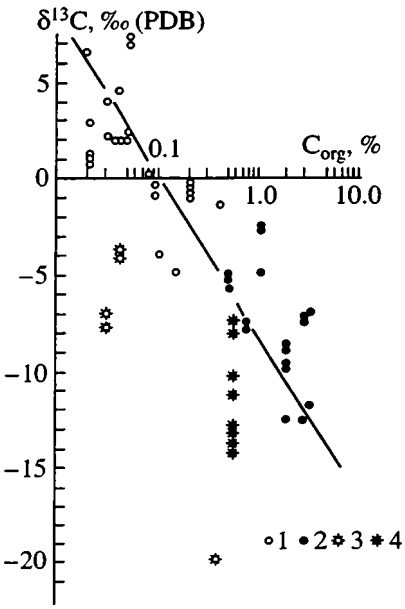

Fig. 4. Correlation diagram of $\delta^{13}C$ in carbonate rocks and diagenetic concretions plotted versus C_{org} in enclosing rocks.

1, 2 – Carbonate rocks; 3, 4 – carbonate concretions; open symbols for C_{org} < 0.5%; closed symbols for C_{org} = 0.5 - 3.5%.

In addition, it was also noted that heavy and superheavy carbon in carbonates ($\delta^{13}C$ > 10‰) originates during sediment diagenesis in both the marine (Murata *et al.*, 1969) and lake (Nissenbaum and Magaritz, 1988; Talbot and Kelts, 1986) basins.

There are several facts that should be taken into account when interpreting data on the Pechenga–Varzuga belt: (1) the time interval under consideration is marked by the most intense dolomite accumulation recognized in the history of the Baltic shield; (2) in addition to carbonates, the Jatulian formations comprise abundant red deposits, which contain totally oxidized iron and associate with borate- and sulfate-bearing beds of a distinctly evaporite origin (Melezhik, 1987); and (3) the Jatulian sedimentation in the Baltic shield was mostly concentrated in shallow continental basins of the protoevaporite type (Melezhik, 1987, 1991; Melezhik *et al.*, 1988). These basins are comparable to the closed, usually saline lakes and shallow marine lagoons that are now accumulating carbonates with abnormally high $\delta^{13}C$ (Botz *et al.*, 1988; Casanova and Hillaire-Marcel, 1987; Talbot, 1990). Examples that are remarkable in this respect are provided by the lakes of the East African rift zone. For instance, the Natron–Magadi lake system, with a total area of more than 20 000 km^2, contains recently formed stromatolites that display $\delta^{13}C$ variations within the range 2.6 - 6.0‰ (Casanova and Hillaire-Marcel, 1987). Carbonates from the Kivu Lake of the same zone have even higher (up to 13.6‰) values of $\delta^{13}C$ (Botz *et al.*, 1988).

As was stated at the beginning of this paper, abnormally heavy carbon in carbonates has also been reported from other regions beyond the Baltic shield. However, it is difficult to prove that all the respective formations are synchronous. Moreover, there are thick carbonate sequences of the Lower Proterozoic that lack notable isotopic anomalies, for instance, the very representative Transvaal Supergroup in South Africa (Shidlowski *et al.*, 1975, 1983) and the Udokan Group in eastern Siberia (Pokrovskii and Melezhik, 1992). Therefore, there is no basis for speculating about a global and long-lasting period of the Early Proterozoic that

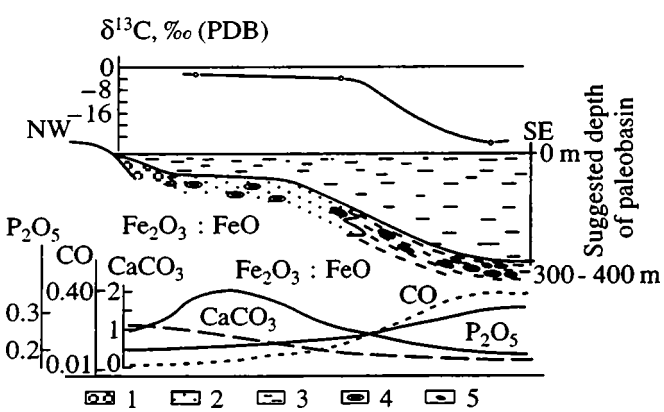

Fig. 5. Variations of $\delta^{13}C$ in diagenetic carbonate concretions and geochemical parameters of enclosing rocks along the inferred facial profile.

1 – Delta deposits; 2 – shallow-water sandstones with dispersed carbonate and hematite; 3 – carbonaceous siltstones of relatively deep origin; 4 – diagenetic calcareous concretions; 5 – diagenetic phosphate concretions.

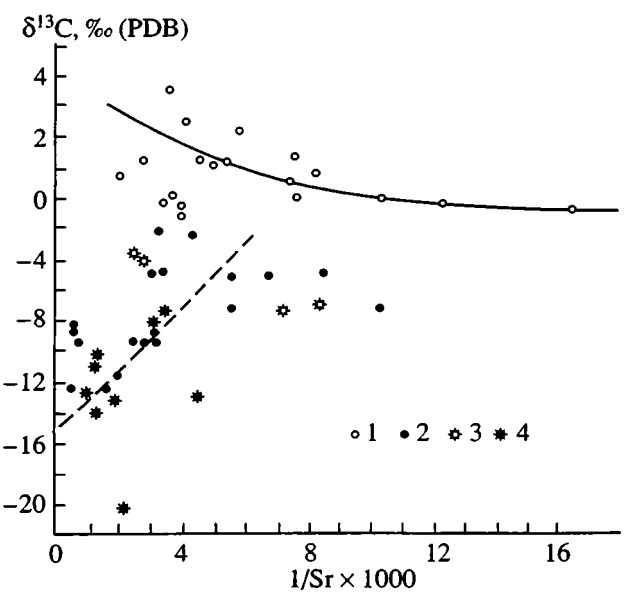

Fig. 6. Correlation diagram of $\delta^{13}C$ versus 1/Sr × 1000 for carbonates enclosed in deposits with two levels of C_{org} concentrations (see Fig. 4 for symbol explanations).

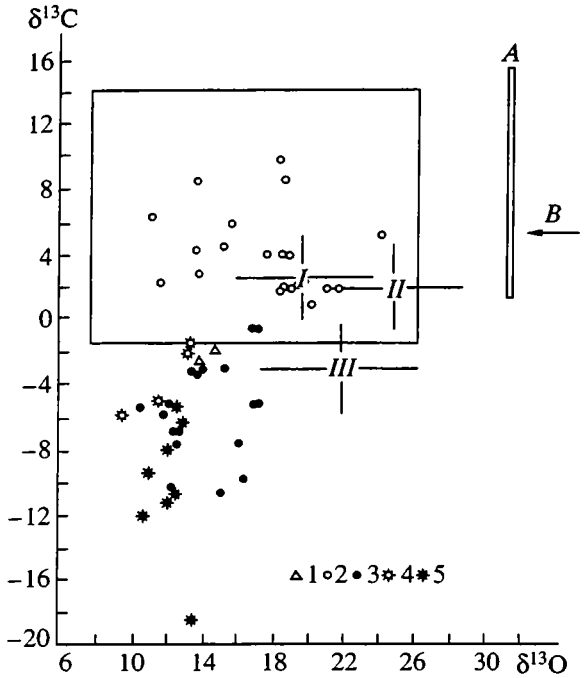

Fig. 7. Correlation diagram of $\delta^{13}C$ versus $\delta^{18}O$ in carbonates of the Pechenga–Varzuga belt.

1 – Sumian carbonate rocks; 2 – Jatulian carbonates from carbon-free deposits; 3 – some Jatulian (three samples) and post-Jatulian carbonates from carbonaceous deposits; 4 – diagenetic carbonate concretions from carbon-free deposits; 5 – similar concretions from carbonaceous deposits.

Rectangular field shows the general diapason of isotopic variations in the Precambrian sedimentary carbonates (Shidlowski *et al.*, 1975, 1983). Crosses indicate: *I* – average (96%) composition of the Precambrian carbonates (Shidlowski *et al.*, 1975); *II* – average composition of marine sedimentary carbonates of the Phanerozoic (Keith and Weber, 1964); *III* – average composition of freshwater sedimentary carbonates of the Phanerozoic (Keith and Weber, 1964). *A* – diapason of $\delta^{13}C$ variations in carbonates from isolated saline lakes (Botz *et al.*, 1988; Casanova and Hillaire-Marcel, 1987; Talbot, 1990); *B* – value for carbonates equilibrated with modern atmosphere $\delta^{13}C$ = –7‰ under 25°C.

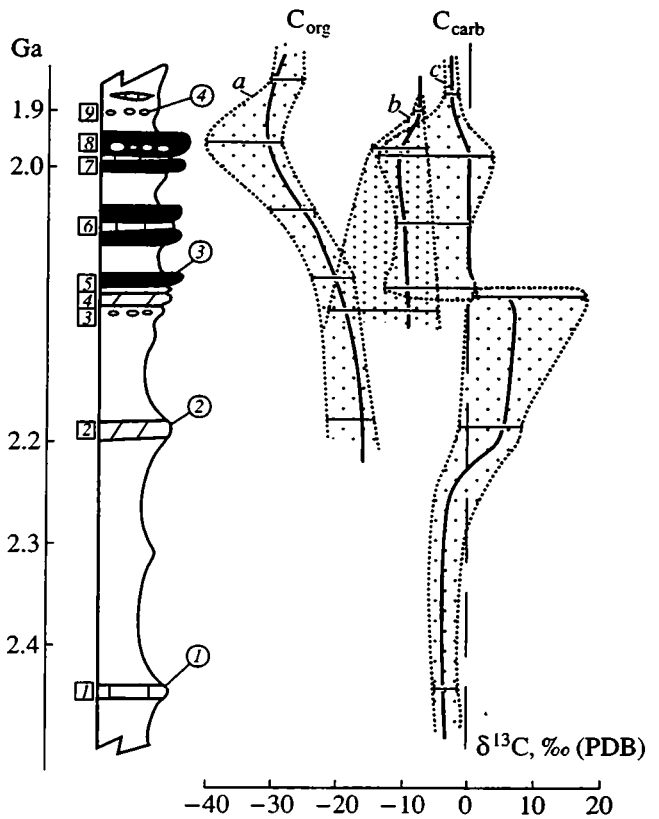

Fig. 8. Variation curves for $\delta^{13}C_{org}$ and $\delta^{13}C_{carb}$ values and position of main carbonaceous and carbonate units in the generalized section of the Lower Proterozoic complex of the Baltic shield.

Numbers in circles for: *1* – limestones; *2* – dolomites; *3* – carbonaceous units; *4* – diagenetic calcareous concretions.

Numbers in squares for the following levels: *1* – Sumian; *2* – lower Jatulian; *3* - *5* – upper Jatulian; *6* - *9* – Lyudy.

Variation curves and diapasons: (*a*) for $\delta^{13}C$ in C_{org} of carbonaceous deposits according to data after Lebedev *et al.* (1987), Karhu (1990), and L.P. Galdobina (personal communication); (*b*) for $\delta^{13}C$ in C_{carb} of diagenetic concretion; (*c*) for $\delta^{13}C$ in C_{carb} of sedimentary carbonates.

resulted in the accumulation of heavy carbon in carbonates.

In this respect, it is also important that, even within the Baltic shield, the accumulation of heavy carbon in carbonates was not strictly synchronous. The Karelian anomaly is of the Late Jatulian age (2.1 - 2.0 Ga; Karhu, 1993), whereas the Kola anomaly corresponds to the interval 2.33 - 2.15 Ga, i.e., to the Early Jatulian time. In contrast, the Upper Jatulian (2.15 - 2.0 Ga) carbonates from the Kola Peninsula show an isotopic composition of carbon that is either normal or alleviated (Fig. 3).

The Upper and post-Jatulian (Lyudy Superhorizon) complexes of limestones and dolomites revealed the most variable values of $\delta^{13}C$ and $\delta^{18}O$. Both parameters are positively correlated, and this suggests that their dispersion was mostly caused by postsedimentary transformations of the rocks. The rocks with the highest values of $\delta^{13}C$ (0 - 3‰) and $\delta^{18}O$ (19 - 24‰) seem to be the least altered, and their C-isotope composition approximates the initial parameter characteristic of sedimentation processes of that time. In Melezhik's opinion, the isotopically heavy carbonates from the Onega zone ($\delta^{13}C$ up to 7.9‰) are the Jatulian in age according to their lithologic features and stratigraphic position in the section; previously, however, they were assigned to the post-Jatulian level (Yudovich *et al.*, 1990).

Carbon dioxide that originated during the decomposition (oxidation) of organic matter undoubtedly contributed much to the formation of rocks with low values of $\delta^{13}C$. Therefore, it seems to be quite symptomatic that these rocks occur in association with beds rich in organic substance (Fig. 4). However, the isotopically light carbonates can be subdivided into two groups:

(1) calcareous concretions and (2) limestones without petrographic indicators of secondary alterations. Each of the groups seems to have been formed under essentially different conditions, because they are dissimilar in their Sr concentrations (moderate in concretions and usually high in limestones) and values of $\delta^{18}O$ (elevated in the first group and low in the second group).

There is no doubt that the concretions were formed during the diagenetic stage. Similar calcareous segregations are widespread in terrigenous and volcano-sedimentary rocks of different ages that are rich in organic matter (Murata, 1969).

The diagenetic model is unsuitable for explaining the origin of the thick limestone beds enriched in the light carbon isotope. In some of them, nearly half of the carbon is of organic ancestry. To oxidize the respective amount of organic carbon, it is necessary that an active oxidizing medium is present, which is not typical of diagenesis. In addition, the diagenetic stage usually results in Sr-depletion, but not in Sr-enrichment of sedimentary carbonates (Brand and Veizer, 1980).

High Sr concentrations, sometimes up to 8000 - 10 000 ppm, are characteristic of brines that originated during the catagenetic stage due to the evaporite deposits being leached by surface waters (Antsiferov, 1989). Accordingly, it is known that high-Sr carbonates rich in the light carbon isotope are either formed because of an interaction of such brines with highly carbonaceous rocks, or else they are formed in the contact zones between oil-bearing horizons and evaporite beds (Sannders *et al.*, 1988; Werner *et al.*, 1988). This suggests that similar catagenetic processes could also be responsible for the isotopic and geochemical peculiarities of the post-Jatulian limestones.

Formally, all the above considerations are quite evident; however, there are many sides to the problem, and they need to be better understood. First, it is surprising that the intense interaction of the post-Jatulian carbonates with the isotopically light carbon dioxide did not affect the petrographic features of the rocks, although this process should be associated with a recrystallization of the rocks with an addition of Sr and Mn. Furthermore, there is the widespread opinion that dolomites are the products of either diagenetic or catagenetic substitution; however, in the Pechenga–Varzuga belt, they show the original sedimentation parameters of carbon isotope composition. In contrast, the isotopic (and geochemical) compositions of the limestones of the belt are definitely disturbed, despite the fact that the sedimentary origin of these rocks is usually taken for granted. Nevertheless, it is clear that the accumulation of thick deposits of organic matter was one of the main factors responsible for the intense secondary alterations of the post-Jatulian carbonates. The oxidation of this matter would be impossible without an equivalent amount of an oxidant, and the sulfate-bearing deposits of that time, now eliminated, were the only rocks suitable to play this role.

Abundant sulfides with high $\delta^{34}S$ (Shatskii, 1990) and many other geochemical and lithological data (Akhmedov, 1990; Grinenko *et al.*, 1989) favor the idea that evaporites were formerly widespread in the Baltic shield. In addition to other data, this study suggests that, after the Jatulian, evaporites were also abundant during the late Lyudy (and post-Lyudy) time. This is directly inferrable from occurrences of catagenetic gypsum beds in carbonaceous deposits of the Transonega region of Karelia (Yu.S. Polekhovskii, personal communication).

CONCLUSION

In the Precambrian history of the Baltic shield, the Early Proterozoic interval was marked by two significant events: the distribution of red deposits at 2.2 Ga and the mass accumulation of organic matter at 2.0 Ga.

Both events were recorded in carbonate rocks as evident anomalies in their carbon isotopic compositions, a positive anomaly for the Jatulian time and a negative one for the post-Jatulian period. However, there are no sufficient reasons to suggest that these anomalies were associated with principal changes in the composition of the Earth's atmosphere, as it is doubtful that these anomalies are synchronous in different regions, for instance, in the Kola Peninsula and Karelian zone. Additionally, the models that have been suggested to relate isotopic effects with the composition of the atmosphere seem to be insufficiently correct.

We believe that the carbon isotope anomalies distinguishable in the Lower Proterozoic deposits of the Kola Peninsula point to the former existence of evaporite formations, which were associated here with rocks rich in organic matter. The anomalies reflect the unusual conditions of past sedimentation and the peculiarities of the postsedimentary processes that are typical of these rock associations. The Lower Proterozoic carbonates, which are rich in the heavy carbon isotope and are abundant in many regions, suggest that, during the respective period, as in some later evolutionary stages (the Vendian-Lower Cambrian, Devonian, and Permian), the conditions on Earth were favorable for the coaccumulation of evaporites and hydrocarbons.

Reviewers V.I. Ustinov and M.A. Semikhatov

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Isotopic Datings of the Lower Cambrian Clays from Estonia and Their Interpretation

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Abstract – Lower Cambrian "blue clays" from the Lontova Formation of northern Estonia, in which previous investigators distinguished three illite generations of different ages, were subjected to a complex study using the methods of X-ray diffraction, electron microscopy, and Rb-Sr and K-Ar isotopic geochronology. The study of the Rb-Sr isotopic system included sample leaching with a solution of ammonium acetate, which was applied to clay fractions with a grain size of < 2.0, 0.6 - 1.0, < 0.6, and 0.1 - 0.4 μm, each containing no less than two illite generations. Despite the linear correlation of data points in the Rb-Sr evolution diagrams, which was revealed for equi-sized fractions of a series of samples, the estimated age values and initial $^{87}Sr/^{86}Sr$ ratios are meaningless in a geological sense. The distinguished relations correspond to mixing lines, suggesting a uniform contribution of different illite generations into equi-sized fractions of clay samples.

Key words: **isotopic dating, Rb-Sr isochron, illite, generations of different age, Lower Cambrian, Estonia.**

INTRODUCTION

Isotopic dating of clay sediments is usually carried out by applying the Rb-Sr and K-Ar methods to whole-rock samples or to fine-sized fractions of clay minerals (Hunziker, 1979; Clauer, 1979, 1982; Gorokhov and Semikhatov, 1984; Pasteels, 1985; Bonhomme, 1987; Semikhatov *et al.*, 1989; Gorokhov *et al.*, 1991). However, clay sediments commonly contain a significant proportion of clastic material, and attempts to date whole-rock samples may only provide approximate and, usually, overestimated age values reflecting the isotopic heterogeneity and presence of inherited Sr and Ar. In addition, Rb-Sr and K-Ar systems of clayey rocks are rarely preserved in a closed state. During subsidence, these rocks suffer a series of diagenetic alterations affecting the potassium- and rubidium-bearing minerals, for instance, the dissolution of K-feldspar and mica followed by illite formation (Hower *et al.*, 1976). As a result, both the rock as a whole and its fine fractions with a grain size of less than 2 μm, which can be used for isotopic dating, may contain more than one generation of clay minerals. This was very clearly demonstrated by Morton (1985), who applied the Rb-Sr method to fractions of the < 0.2 and 1 - 2-μm from Upper Devonian black shales in Texas, USA, and estimated their isochron ages as 302 and 539 Ma, respectively. The first value was interpreted as the age of diagenesis responsible for the origin of the authigenic 1Md illite, and the second one was interpreted as the age of detrital 2M illite characterizing the source material of the Upper Devonian shales.

The applicability of one or another sedimentary rock or its fraction for dating diagenetic processes is usually estimated by measuring (in Δ° 2θ units) the illite crystallinity index (I_k), using the width of the 10 Å peak at its half-height level on the diffractogram of oriented samples (Kubler, 1966). According to the recommendations of the Conference on Diagenesis and Metamorphism held in Bristol, 1984, the boundary between the epi- and the anchizone is at the level of 0.25°, and the next boundary, separating the anchizone and the zone of diagenesis, is at the level of 0.42°. If, however, the material under study contains several illite generations with different crystallinity indexes, the measured I_k value may represent a random result of weighted summing of these individual characteristics (Reuter, 1985). Therefore, when Rb-Sr and K-Ar methods of age determinations are applied to whole-rock samples or to clay fractions with a significant range in grain sizes, there is a chance that such mixing parameters will be mistaken for the age markers of "real events," although they are meaningless with respect to the actual geological and geochemical processes. This side of the problem is still not given sufficient consideration, especially in cases where Rb-Sr data points show a linear correlation in the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagram. Provided that this correlation is coupled with "appropriate" I_k values and a "reasonable" level of the initial $^{87}Rb/^{86}Sr$ ratio approximating this parameter in the sedimentation medium (the sea water), some researchers erroneously consider the obtained age determinations to be well-constrained (Zhang *et al.*, 1984; Krasnobaev *et al.*, 1986).

The objectives of this work included a detailed geochronological study of sedimentary rocks containing more than one generation of clay minerals. We used standard procedures and the technique of Rb-Sr and

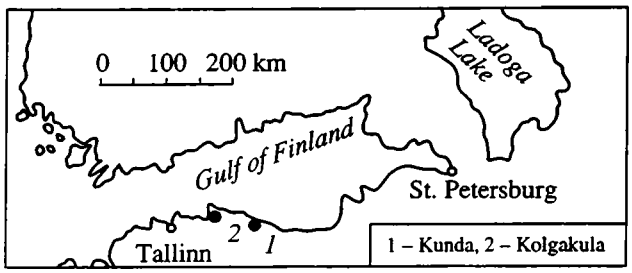

Fig. 1. Geographic scheme of sampling localities.

K-Ar dating, and we determined the Lower Cambrian blue clays from northern Estonia to be suitable for our goals.

GEOLOGICAL POSITION AND PREVIOUS STUDIES OF THE BLUE CLAYS

Blue clays of the Lontova Formation of the Baltic Group are exposed along the southern coast of the Gulf of Finland. The stratotype section of this formation (Mens and Pirrus, 1977) was used as a basis to distinguish the pretrilobite part of the Cambrian within the entire East European platform (*Stratigrafiya ...*, 1979). Recent data (Bowring *et al.*, 1993) show that the age of the Lontova sediments is ca. 530 Ma.

The studied samples were collected near the cities of Kunda (the coastal quarry of a cement factory, sample series 806) and Loksa (the Kolgakula quarry, sample series 807). They represent homogenous compact marine clays of the Kestla Member of the Lontova Formation (Fig. 1); in both localities, they are typically variable in color from violet-brown to greenish-gray. The main component of the rocks is dioctahedral illite. In fine fractions, it seems to be a constituent of the mixed-layer minerals (illite-smectite), with a low proportion of expanding layers.

Petrographic study of thin sections (Reyer, 1967) revealed that 10 - 30% of the clay minerals in the rock matrix are authigenic in origin, and they seem to have been formed by colloid coagulation. In addition, the rocks showed many signs of the transformation of terrigenous components into illite. Apart from micas, which suffered the most intense hydration in their pelitic and fine-aleuritic fractions, this is also true for feldspar and quartz grains, because there are reaction rims of authigenic illite surrounding these grains. As for the feldspar, illite also appears attached to its cleavage faces or as complete pseudomorphs after feldspar grains. The preservation of such pseudomorphs suggests that they were formed under the quiet conditions of diagenesis, because any redeposition would have destroyed them.

Using transmission electron microscopy, Firsov *et al.* (1971) showed that fractions of the Lontova clays with a grain size of <5 μm contain illite consisting of equant and elongated crystallite plates; the latter were suggested to be the alteration products of the equant crystallites.

Further X-ray, electron-microscopic, and Rb-Sr studies on the Lontova clays (Gorokhov *et al.*, 1982, 1994) revealed that they bear three illite generations: one was identified as high-T detrital (fraction size of 0.6 - 2 μm), and the two others were identified as low-T authigenic, concentrated in fractions with a grain size of 0.1 - 0.2 and <0.1 μm. The revealed trend of older illite preservation in coarse fractions and the appearance of younger illite generations in fine fractions are easily understandable because of the diagenetic and catagenetic recrystallization ability of fine particles. The established age difference between the early and late generation is ca. 250 Ma.

METHODS

Each of the studied samples was approximately 1 kg in weight. Powdered samples were suspended in distilled water to obtain their fractions using the classical sedimentation method. Fractions for analysis had size ranges of <2, 0.6 - 1, and <0.6 μm. Other fractions with narrow ranges in grain size, including that of 0.1 - 0.4 μm, were separated for electron-microscopic study by the centrifuging technique (Turchenko *et al.*, 1993). According to Gorokhov *et al.* (1994), fractions with a grain size of 0.6 - 1 and 0.1 - 0.4 μm contain the first and second generations of illite, which were distinguished in the Lontova clays, while coarser fractions (grain sizes of < 2 and <0.6 μm) incorporate all three of the known generations of this mineral.

The mineralogical compositions of the fractions and the crystallinity index of the illite were determined using the X-ray diffractometer DRON-2.0. To obtain desirable specimens, the aliquot of each fraction dispersed in distilled water was precipitated on a glass plate. The X-ray scans were run at a rate of 1° 2θ/min from 4° to 55° under CoK_α radiation. The procedure was repeated after saturating the samples with ethylene glycol.

Transmission electron microscopes JEM-100B and Philips EM-300 were used to study the morphology of the clay particles, which were first precipitated from the fraction aliquots suspended by ultrasonic treatment and then dried on a carbon-covered grid. Respective images were obtained at a magnification diapason of 8000 - 50 000.

We used the isotope dilution method to measure Rb and Sr concentrations and $^{87}Sr/^{86}Sr$ ratios after dissolving the fractions and extracting the elements by the standard ion-exchange technique. The details of this procedure are described by Varshavskaya *et al.* (1975). The isotopic composition of Rb and Sr was determined using the mass spectrometers MI-1201 and MI-1320, respectively. An average value of 0.70800 ± 0.00009 was determined for the Eimer and Amend standard during the period of data measurement (in this paper, all

uncertainties are quoted at the 2σ level). The analytical uncertainty for the $^{87}Rb/^{86}Sr$ ratio in geological samples was estimated to be ±2%, and that for the $^{87}Sr/^{86}Sr$ ratio was estimated to be ±0.1%. The regression approach of McIntyre *et al.* (1966) was used to determine the slope, intercept, and mean square of weighted deviates (MSWD). It was modified by assuming a constant percent error for the $^{87}Sr/^{86}Sr$ ratio and by applying Williamson's method for calculating the regression parameters (Williamson, 1968; Shukolyukov *et al.*, 1974). The precision and accuracy of the technique were controlled by analyses of NBS 70a K-feldspar with average Rb and Sr contents of 525 and 66.0 μg g^{-1}, respectively, which were measured during our working period. Leaching with 1N solution of ammonium acetate (NH_4OAc) was used to analyze the Sr incorporated into the structure of the clay minerals separately from the easily soluble Sr, e.g., from that adsorbed by the outer faces of the particles (Morton and Long, 1980; Morton, 1985).

We used flame photometry to measure K contents in the samples. The Ar measurements were done by the isotope dilution technique with a ^{38}Ar tracer using a complex of equipment consisting of the following: a gas extraction system; the mass spectrometer MI-1330 for Ar isotope analysis; and the computer Elektronika D3-28, programmed to control the measurement process and data processing (Drubetskoi and Sprintsson, 1982). Biotite A70 was used as a standard. The analytical uncertainty is assumed to be ±0.2% for the $^{40}Ar/^{38}Ar$ ratio and ±2 - 4% for the $^{40}Ar/^{38}Ar$ ratio measurements. According to the data of duplicate determinations, the estimated precision is ±3% for both the K and the radiogenic Ar concentrations. The used samples, 0.05 g in weight, were not preheated before analyses.

The decay constants recommended by the International Subcommission on Geochronology (Steiger and Jager, 1977) were used to calculate respective age values.

RESULTS

Table 1 shows the estimated percentages of different fractions in a total mass of clay particles which are less than 2 μm in size. The mineral composition of studied fractions and the illite crystallinity index are presented in Table 2. Fractions with a grain size of 0.6 - 1 and 0.1 - 0.4 μm mostly consist of illite (80 - 90%) and chlorite (10 - 20%), with a minor admixture of kaolinite and quartz. Fractions with a grain size of < 0.6 μm contain illite, mixed-layer illite-smectite with an insignificant amount of expanding layers (95% in total), and chlorite (up to 5%). For particles that are 0.6 - 1 μm in size, the illite crystallinity index varies from 0.25°, the value typical of the anchizone, to 0.60°, which is characteristic of the diagenesis zone. In fractions with particle sizes of 0.1 - 0.4 μm, respective values are more homogenous (from 0.40° to 0.60°), approaching the boundary value (0.42°) which separates the zones of anchimetamorphism and diagenesis. If particle sizes are <0.6 μm, the range of the crystallinity index is 0.80° - 1.40°. These values are typical of nonmetamorphic mixed-layer illite-smectite; therefore, according to the usual criteria, one could suspect that the respective fractions are suitable for dating the stage of diagenesis in the Lontova sediments.

Table 1. Weight percentage of size-ranked particles in clay fractions <2 μm from the Lontova Formation

Sample number	Particle sizes, μm			
	<0.1	0.1 - 0.4	0.4 - 0.6	0.6 - 2
Kunda quarry				
806/1	34	51	8	7
806/8	31	54	7	8
806/9	38	57	4	1
Kolgakula quarry				
807/6	32	57	7	4
807/8	27	58	7	8
807/11	31	55	7	7
807/13	32	56	9	3

The results of our electron-microscopic study of the clay fractions are illustrated in Fig. 2. An irregular form of the particles is typical of the coarse fractions (Fig. 2a), while in finer fractions, they are lathlike, represented by tolerably regular elongated plates (Figs. 2b and 2c). In addition, in coarse and transitional fractions, there are notable reaction rims of fine plates surrounding the irregular grains (Figs. 2a and 2b).

The data of Rb-Sr analyses (Table 3) show that the data points corresponding to 17 fractions with grain sizes of <0.6 μm plot in the Rb-Sr evolution diagram along an isochron (Model 1, McIntyre *et al.*, 1966), which has a slope corresponding to an age of 556 ± 43 Ma, an initial ratio of $^{87}Sr/^{86}Sr = 0.7096 \pm 0.0036$, and MSWD = 1.4. The regression line for these and additional data points, which represent the leaching products of similar fractions <0.6 μm from samples 807/8 and 806/9, corresponds to an errorchron (Model 2, McIntyre *et al.*, 1966) with MSWD = 2.6. The estimated age of the errorchron is 543 ± 6 Ma, and the initial ratio of $^{87}Sr/^{86}Sr = 0.7110 \pm 0.0004$ (Fig. 4). As was stated earlier (Gorokhov *et al.*, 1991), the statistic models of McIntyre *et al.* (1966) become meaningless in geochemical aspect when they are used to process data obtained using the leaching method. We use them here by estimating the ages and initial $^{87}Sr/^{86}Sr$ ratios to account only for the scatter in data points exceeding that attributable to analytical uncertainty. Data for fractions with a grain size of 0.6 - 1 μm and for the products of their leaching resulted in an errorchron (Model 4, McIntyre *et al.*, 1966; MSWD = 4.8) with an age value

Table 2. Mineralogy of clay fractions from the Lontova Formation

Sample number	Particle sizes, μm	Content of minerals, %				Crystallinity index of illite, Δ°2θ
		mixed-layer illite-smectite	chlorite	kaolinite	quartz	
806/1	<0.6	95	5	–	–	1.00
	0.1 - 0.4	90	10	–	n. d.	0.50
806/2	0.6 - 1	80*	20	n. d.	n. d.	0.25
	0.1 - 0.4	90	10	–	n. d.	0.40
	<0.6	95	5	–	–	1.35
806/3	<0.6	95	5	–	–	1.00
806/5	<0.6	95	5	–	–	0.90
806/6	<0.6	95	5	–	–	1.20
806/7	<0.6	95	5	–	–	1.40
806/8	<0.6	95	5	–	–	0.80
	0.1 - 0.4	85	15	–	n. d.	0.40
806/9	0.6 - 1	80*	20	n. d.	n. d.	0.25
	<0.6	95	5	–	–	1.20
	0.1 - 0.4	90	10	–	n. d.	0.55
806/10	<0.6	95	5	–	–	1.10
806/17	<0.6	95	5	–	–	1.40
807/5	<0.6	95	5	–	–	1.10
807/6	<0.6	95	5	–	–	1.10
	0.1 - 0.4	85	15	–	n. d.	0.50
807/8	0.6 - 1	90*	10	n. d.	n. d.	0.60
	<0.6	95	5	–	–	1.00
	0.1 - 0.4	80	20	–	n. d.	0.60
807/9	<0.6	95	5	–	–	1.40
807/10	<0.6	95	5	–	–	1.00
807/11	0.6 - 1	80*	20	n. d.	n. d.	0.25
	<0.6	95	5	–	–	1.30
	0.1 - 0.4	85	10	–	n. d.	0.50
807/12	<0.6	95	5	–	–	0.90
807/13	0.6 - 1	85*	15	n. d.	n. d.	0.50
	<0.6	95	5	–	–	0.90
	0.1 - 0.4	90	10	–	n. d.	0.50

* Illite.

of 612 ± 7 Ma, initial $^{87}Sr/^{86}Sr = 0.7110 \pm 0.0004$, and MSWD = 4.8 (Fig. 5). A comparable age (593 ± 7 Ma) and initial $^{87}Sr/^{86}Sr$ ratio (0.7124 ± 0.0004) were obtained for the isochron (Model 1, MSWD = 0.4), characterizing fractions with particles that are 0.1 - 0.4 μm in size and the products of their leaching (Fig. 6).

The K-Ar data for fractions with grain sizes of <2, <0.6, and 0.1 - 0.4 μm are shown in Table 4. The apparent age is in the range of 500 - 556 Ma for the coarse fractions (<2 μm). For finer fractions, it is either 475 - 522 Ma (grain size <0.6 μm), or 488 - 498 Ma (grain size 0.1 - 0.4 μm).

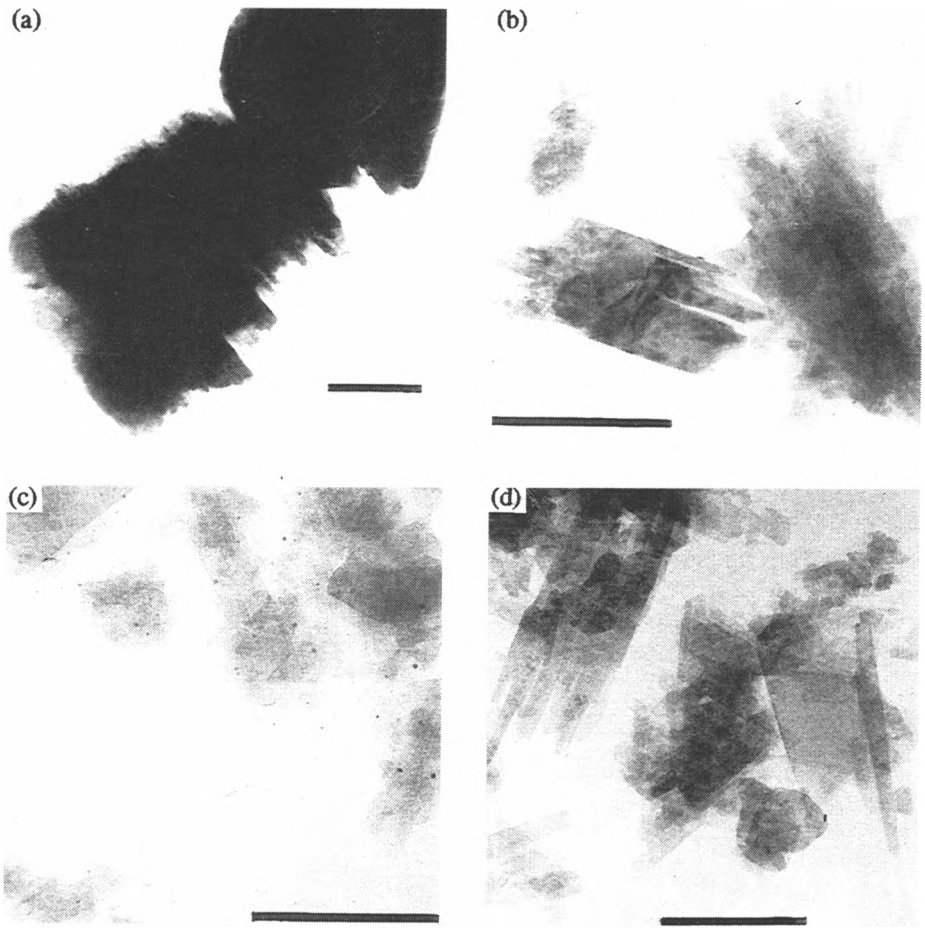

Fig. 2. Electron-microscopic images of mineral particles from different grain-size fractions of sample 806/8: (a) fraction 0.6 - 2 μm (Gorokhov *et al.*, 1994), bar – 1 μm; (b) fraction 0.3 - 0.4 μm, bar – 0.5 μm; (c) fraction 0.1 - 0.2 μm, bar – 0.5 μm; (d) fraction <0.1 μm, bar – 0.5 μm.

DISCUSSION OF RESULTS

The results of X-ray diffraction, electron microscopy, and isotopic geochronology (Reyer, 1965; Firsov *et al.*, 1971; Gorokhov *et al.*, 1982, 1989, 1994) show that fractions of the Lontova clays with a grain size of <2 μm contain three illite generations. The first generation is represented by detrital 2M illite with low I_k (ca. 0.20°), which is characteristic of the epi- or anchizone (Fig. 7). In photographs, which were obtained using transmission electron microscopy, this illite in fractions with a grain size of 0.6 - 2 μm consists of isometric plates with irregular margins (Fig. 2a). The minimum age of this generation is 722 ± 13 Ma (Fig. 8). The clay fractions with a grain size of 0.1 - 0.2 μm include authigenic 1M illite of the second generation, which was formed 533 ± 8 Ma ago (Fig. 8) during the burial and diagenesis of sediments. This illite with a high crystallinity index (I_k = 1.05° - 1.25°) featuring the diagenesis zone (Fig. 7) is represented by lathlike particles (Fig. 2b). Comparable mineralogical characteristics (I_k = 1.10° - 1.80°, Fig. 7) and morphologic features (Fig. 2d) were recognized in the authigenic 1M(1Md) illite of the third generation, concentrated in fractions with a grain size of <0.1 μm. Many mineral particles also show here bladed forms, pointing to their crystallization under the conditions of a high water/rock ratio (Kotel'nikov *et al.*, 1992). The last illite generation is thought to have been formed 470 - 480 Ma ago, when ascending movements transposed sediments into the circulation zone of meteoric waters.

It is remarkable that we were able to attain a reasonably complete separation for only the second generation of illite concentrated in fractions with a grain size range of 0.1 - 0.2 μm. Other fractions with particle sizes of 0.2 - 2 μm and <0.1 μm definitely contain mixtures of the first-second and second-third illite generations, respectively. This is evident from data points characterizing pertinent residues after leaching, because in the Rb-Sr evolution diagram, they are confined to the mixing line of pure components (Gorokhov *et al.*, 1994). In addition, the microphotographs clearly illustrate that individual illite laths or fibers grow around cores of detrital grains available in coarse fractions (Figs. 2a and 2b).

Table 3. Data of Rb-Sr analysis

Sample number	Particle sizes, μm	Analyzed material*	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
806/1	<0.6	US	202	107	5.474	0.75234
			203	107	5.497	0.75287
806/2	0.6 - 1	US	227	82.3	8.046	0.77823
		L	1.65	15.8	0.3027	0.71372
		R	220	63.8	10.08	0.80104
	<0.6	US	206	101	5.925	0.75708
806/3	<0.6	US	205	107	5.584	0.75302
806/5	<0.6	US	206	89.3	6.716	0.76190
806/6	<0.6	US	209	98.5	6.177	0.75849
806/7	<0.6	US	215	115	5.417	0.75166
806/8	<0.6	US	205	86.0	6.937	0.76500
			206	86.4	6.593	0.76511
	0.1 - 0.4	US	222	80.1	8.100	0.78117
		L	2.79	19.2	0.4215	0.71582
		R	223	61.0	10.68	0.80227
806/9	0.6 - 1	US	225	72.8	9.022	0.79137
		L	1.59	13.1	0.3531	0.71367
		R	227	58.5	11.33	0.81070
	<0.6	US	213	95.7	6.460	0.76046
806/10	<0.6	US	209	93.2	6.526	0.76046
806/17	<0.6	US	212	101	6.095	0.75798
807/5	<0.6	US	211	107	5.738	0.75433
807/6	<0.6	US	210	122	5.006	0.74948
807/8	<0.6	US	211	112	5.497	0.75442
		L	2.52	34.8	0.2102	0.71273
		R	204	71.6	8.282	0.77749
		R**	207	79.8	7.566	0.77099
		R**	205	83.9	7.101	0.76835
	0.1 - 0.4	US	223	97.5	6.659	0.76800
		L	2.51	26.6	0.2737	0.71475
807/9	<0.6	US	211	106	5.789	0.75602
		L	3.03	35.8	0.2451	0.71307
		L**	1.92	24.9	0.2125	0.71283
		R	211	72.4	8.502	0.77678
807/10	<0.6	US	212	109	5.658	0.75434
807/11	0.6 - 1	US	205	76.2	7.824	0.78033
	<0.6	US	213	104	5.956	0.75735
807/12	<0.6	US	210	101	6.058	0.75794
807/13	0.6 - 1	US	239	98.5	7.072	0.77070
		L	2.15	18.8	0.3311	0.71437
	0.1 - 0.4	US	229	91.4	7.293	0.77436
		L	2.19	23.3	0.2717	0.71482
		R	223	65.1	10.01	0.79745

Notes: * US – untreated sample, L – leachate, R – residue.
** Control leaching run.

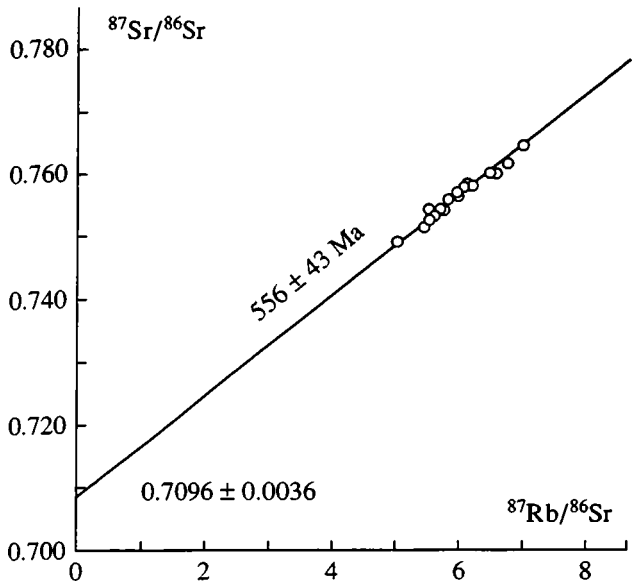

Fig. 3. Rb-Sr evolution diagram for the Lontova clay fractions with grain size <0.6 μm.

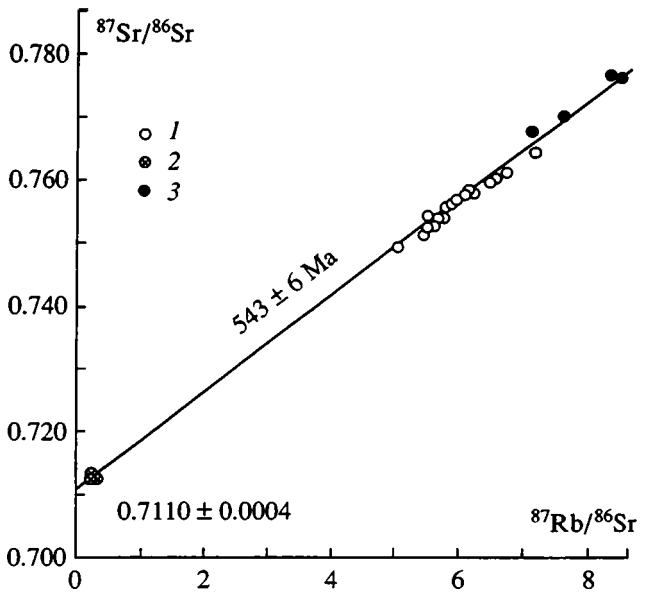

Fig. 4. Rb-Sr evolution diagram for the Lontova clay fractions with grain size <0.6 μm and their leachates after treatment with 1N solution of ammonium acetate.
1 – untreated sample; *2* – leachate; *3* – residue.

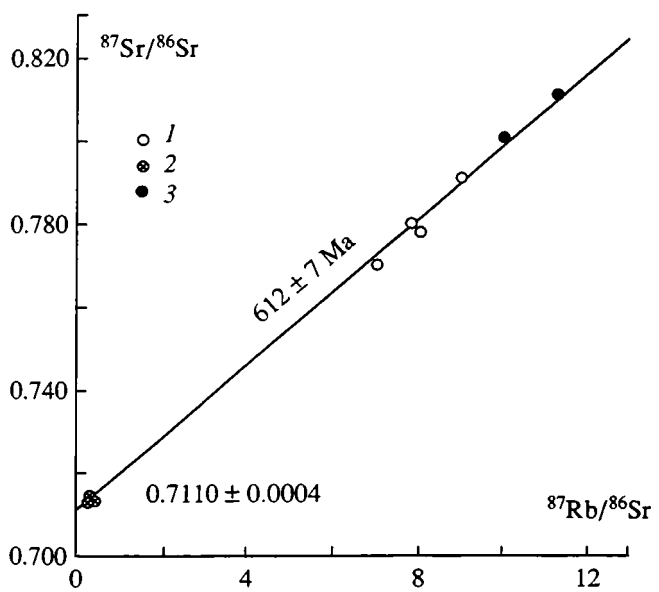

Fig. 5. Rb-Sr evolution diagram for the Lontova clay fractions with grain size of 0.6 - 1 μm and their leachates after treatment with 1N solution of ammonium acetate (see Fig. 4 for symbol explanations).

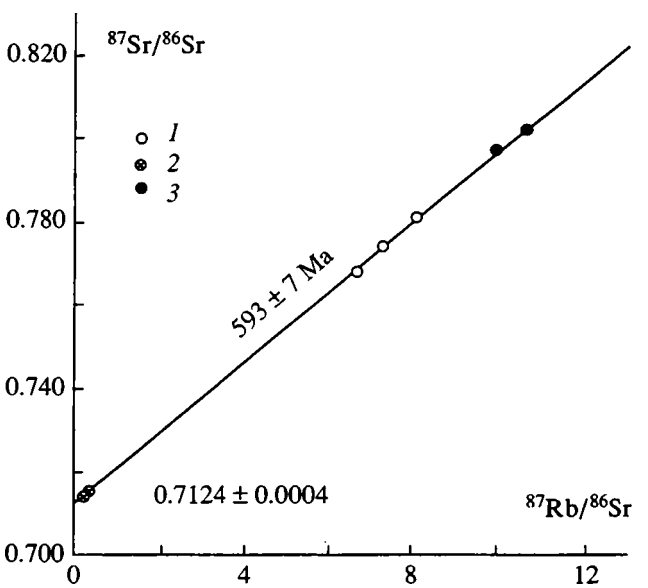

Fig. 6. Rb-Sr evolution diagram for the Lontova clay fractions with grain size of 0.1 - 0.4 μm and their leachates after treatment with 1N solution of ammonium acetate (see Fig. 4 for symbol explanations).

In the Rb-Sr evolution diagrams, the regression lines for groups of data points corresponding to an untreated sample, its acetate leachate, and residue represent, in general, the mixing lines and reflect the balance of the material. They could be considered as isochrons, the slopes of which, are proportional to the ages of the studied mineral phases (or mineral associations). However, the intercept of such an isochron points to the $^{87}Sr/^{86}Sr$ ratio in the mineral-forming medium only if the minerals and interstitial solution, presumably represented by leachate, were under isotopic equilibrium at the moment of crystallization or the diagenetic recrystallization of mineral phases. Such situations for various alumosilicates of low-T origin have been described in several publications (Clauer, 1982; Clauer *et al.*, 1982, 1984, 1990; Kralik, 1984). However, data on

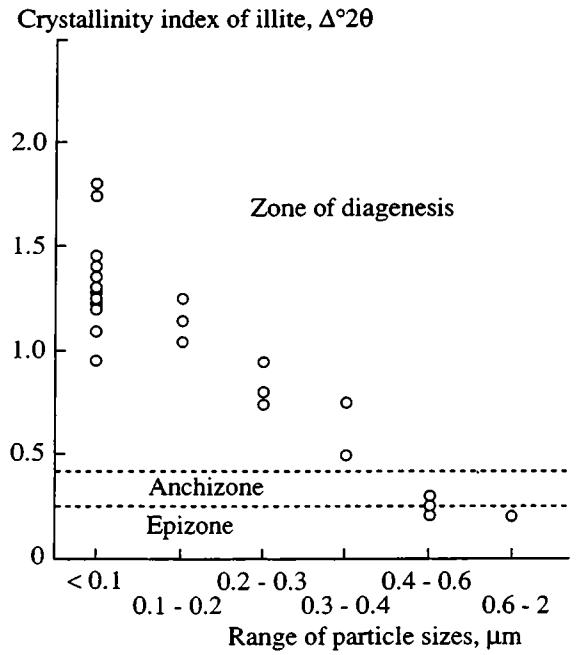

Fig. 7. Variations of illite crystallinity index (I_k) in the Lontova clay fractions of different grain size (after Gorokhov *et al.*, 1994).

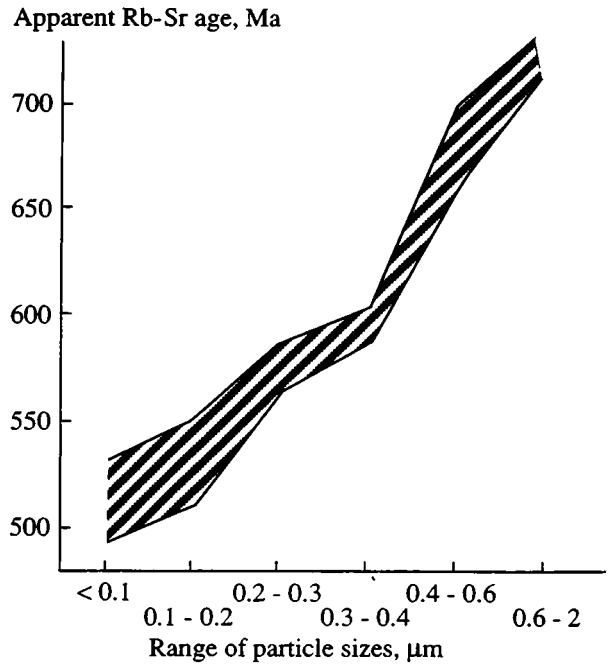

Fig. 8. Diagram of apparent Rb-Sr ages (±2σ) for the Lontova clay minerals plotted versus grain sizes of mineral fractions (after Gorokhov *et al.*, 1994).

the Lower Cambrian clays of the Lontova Formation, which were obtained by leaching their fractions with 1N solution of ammonium acetate (Gorokhov *et al.*, 1989, 1994), show that the three described illite generations were crystallized in surroundings with different Sr isotope compositions (Fig. 9). The initial $^{87}Sr/^{86}Sr$ ratios were 0.7100 ± 0.0006 for the first, 0.7137 ± 0.0010 for the second, and 0.7115 ± 0.0004 for the third generations. It is very significant that none of these values correspond to the $^{87}Sr/^{86}Sr$ ratio in the sea water of those times. Thus, even the illite fraction, which is attributed to the stage of subsidence and diagenesis, was formed in a medium enriched in radiogenic ^{87}Sr extracted by the dissolution of clastic components of the clays (see also Clauer *et al.*, 1982, 1984; Holtzapffel *et al.*, 1985). This casts doubt on the criterion that is widely used to control the correctness of age determinations for clastic rocks, and it implies that $^{87}Sr/^{86}Sr$ ratios should be identical in diagenetic minerals and the sea water surrounding sediments.

Provided that all the analyzed fractions contain two or three illite generations significantly different in their ages, the estimated Rb-Sr and K-Ar ages represent weighted averages for respective mixed minerals, and they do not correspond to real events in the geological history of the rocks. Nevertheless, the data points for these fractions show linear trends in the Rb-Sr evolution diagram, and their deviations from the respective regression lines are at the level of analytical uncertainty (Figs. 3 and 6) or insignificantly higher (Figs. 4 and 5). The estimated values of the initial $^{87}Sr/^{86}Sr$ ratios are also near the range typical of the Cambrian sea water (Burke *et al.*, 1982; Veizer *et al.*, 1983). Without additional information on the mineralogy and geological history of the studied clays, but with due regard for the X-ray data, the linear trends in Figs. 3 - 6 could be considered as satisfying the basic criteria recently used by Rb-Sr dating of sedimentary rocks. The same is true for the K-Ar age determinations (Table 4), the scatter of which seems to be within a reasonable range. In accordance with available recommendations (Clauer, 1979; Bonhomme, 1987), the obtained data could be interpreted in the following manner: 612 Ma (Rb-Sr method, fraction of 0.6 - 1 μm) is the age of metamorphism that affected the source rocks of the detrital illite, 543 - 593 Ma (Rb-Sr method, fractions of 0.1 - 0.4 and < 0.6 μm) is the age of early diagenesis in the Lontova sediments, and 475 - 522 Ma (K-Ar method, fractions of 0.1 - 0.4 and < 0.6 μm) is the age of the final diagenetic stage in the studied rocks. However, this interpretation is misleading.

The meaningless (with respect to geology) linear trends that are observable in the Rb-Sr evolution diagrams for the Lontova clays are a result of the very uniform distribution of heterogeneous material in these sediments. As follows from data on the distribution of different particles in the fractions with a grain size of <2 μm (Table 1), there is a reliable similarity between all of the studied samples, although they characterize two localities that are separated from one another by dozens of kilometers. All the fractions consist mostly of particles that are less than 0.4 μm in size and compose

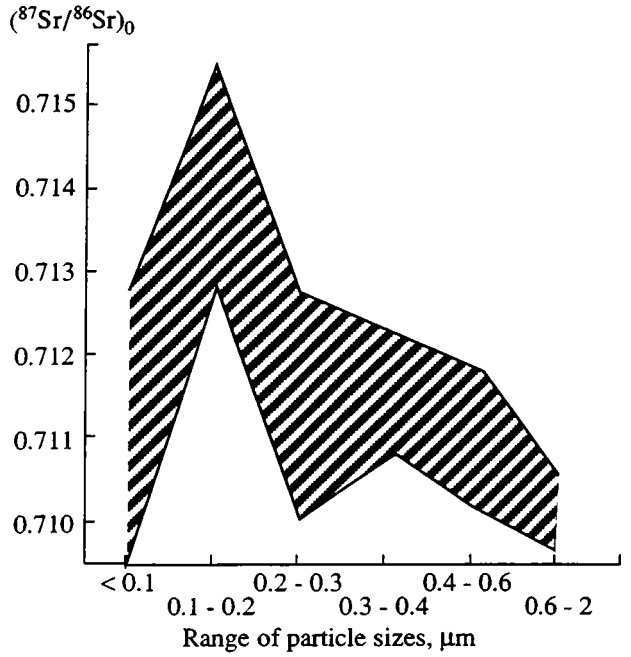

Fig. 9. Diagram of initial $^{87}Sr/^{86}Sr$ ratios (±2σ) in the Lontova clay minerals plotted versus grain sizes of mineral fractions (after Gorokhov *et al.*, 1994).

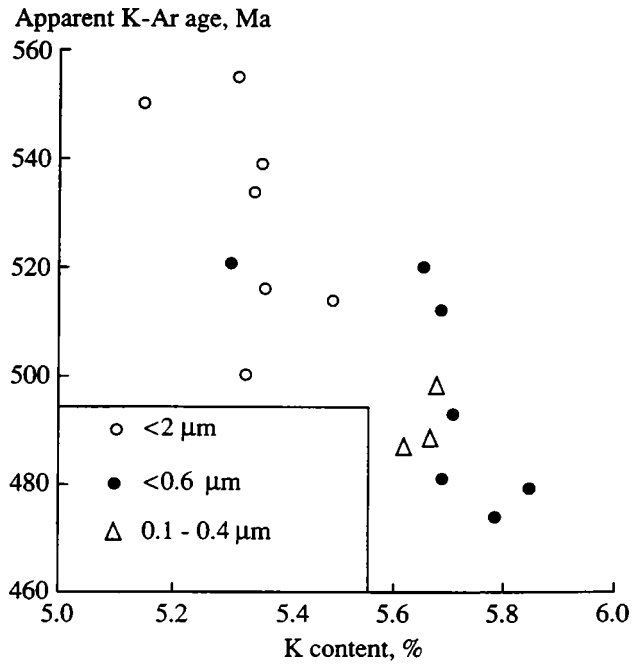

Fig. 10. Diagram of apparent K-Ar ages for the Lontova clay fractions plotted versus their K content.

either 27 - 38% (grain size <0.1 μm) or 51 - 58% (grain size 0.1 - 0.4 μm) of the fraction volume. The proportion of particles ranging in dimension between 0.4 and 2 μm is less than 15%. Given such a distribution of particles and stable fraction sorting by the sedimentation method, an equi-sized fraction from different samples should contain similar proportions of different mineral generations. As a result, we should obtain in the Rb-Sr evolution diagram linear regressions of data points with similar values of an "averaged age." The other consequence of the "uniform heterogeneity" of the Lontova clays will be the similarity of their equi-sized fractions with respect to averaged initial ratios of $^{87}Sr/^{86}Sr$, despite the fact that these ratios are different in various illite generations.

With due regard for the data presented in Table 1, we may find a reasonable explanation for the distinguished difference in Rb-Sr ages of various fractions (Figs. 3 - 6). The fraction with a grain size of <0.6 μm contains a significant portion of particles that are less than 0.1 μm in size. These particles are represented mostly by the third illite generation, dated at 470 - 480 Ma (Gorokhov *et al.*, 1994). Evidently, this is a good reason to explain why this fraction revealed a younger "average" value for its Rb-Sr age as compared to fractions with particle sizes of 0.6 - 1 and 0.1 - 0.4 μm, which are rich in older illite of the first and second generations.

The mixing systematic is also evident from the K-Ar ages measured for different fractions. Figure 10, in which the K-Ar age values are plotted versus K content in the analyzed fractions, shows a distinct negative correlation of both parameters. As may be expected, the oldest values of apparent age are typical here of coarse fractions with particles of <2 μm, among which the first illite generation is relatively abundant. Fractions with particles of <0.6 and 0.1 - 0.4 μm have younger apparent ages because of the higher proportion of authigenic illite in their composition. According to Gorokhov *et al.* (1994), the higher K contents in the last two fractions are correlative with some increase in Rb concentrations in the authigenic illite of the second generation (grain size 0.1 - 0.2 μm) as compared to its first generation (grain size 0.6 - 2 μm). This indicates that K and Rb enter the structure of newly formed authigenic illite when these elements become extracted from clastic grains of feldspar and mica during the subsidence and diagenesis of sediments. Provided that none of fractions that were analyzed in this study are represented by any pure or highly concentrated illite generation, all the K-Ar ages in Table 4 and Fig. 10 are meaningless in relation to geology.

CONCLUSION

The Rb-Sr and K-Ar analyses of clay fractions that have a wide diapason of mineral particles may result in isotopic dating that is meaningless in a geological sense if these fractions contain more than one generation of illite. The illite crystallinity index is useless as a criterion of illite origin, because its value depends on the proportions of different generations of this mineral and

Table 4. Data of K-Ar analysis

Sample number	Particle sizes, μm	K, %	$^{40}Ar_p$, ppb	K-Ar age, Ma
806/8	0.1 - 0.4	5.62	218	488
806/17	<2	5.13	230	551
		5.17		
	<0.6	5.31	223	522
807/5	<2	5.37	223	517
	<0.6	5.79	218	475
807/6	<2	5.33	213	500
	<0.6	5.71	225	494
807/8	0.1 - 0.4	5.67	220	488
807/9	<2	5.35	231	535
	<0.6	5.66	237	521
807/11	<2	5.36	235	540
			233	
	<0.6	5.69	218	482
807/12	<2	5.32	240	556
	<0.6	5.69	234	513
807/13	<2	5.49	227	515
	<0.6	5.85	223	480
	0.1 - 0.4	5.68	226	498

has no reliable meaning. Provided that stable sample fractionation occurs after treatment by the sedimentation method, the data points for equi-sized fractions from different samples may define linear regression lines in the Rb-Sr evolution diagram, the age of which is meaningless, even if their intercepts indicate initial ratios typical of the Sr isotope composition in the sea water of the respective geological epoch.

To minimize probability of a misleading interpretation of such linear Rb-Sr evolutionary trends and correlative K-Ar data, we recommend the following:

(1) Clay fractions for analysis should be as uniform in grain size as possible. This decreases the chance that the fractions will contain more than one illite generation.

(2) Additional data from X-ray study, electron microscopy, and isotopic geochronology for several fractions that are different in grain size will be useful for detecting the possible coexistence of various illite generations.

(3) In combination with analytical results obtained by fraction leaching with ammonium acetate, the Rb-Sr method provides important information on the initial Sr isotope composition in illite, and it enables those fractions that contain the end-members of mixed systems to be identified.

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Regional Middle-Upper Cambrian Stratigraphic Scheme of the East European Platform

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Abstract – A regional stratigraphic scheme of Middle-Upper Cambrian deposits is elaborated. The Middle Cambrian comprises the Kibartai, Veselovsk, and Lukov horizons and the Upper Cambrian incorporates the Tolbukhino, Vorchin, Tsitre, Volodarsk, and Ladoga horizons. All of the units are substantiated by acritarchs and, except for the Kibartai Horizon, established for the first time. Paleontological characteristics of the horizons include an indication of faunal and acritarch reference taxa. No regional horizon is recognized within the Deimena deposits and their analogues, which are situated between the Kibartai and Veselovsk horizons, because of insufficient data. A continuous succession of all the horizons in a single section was not observed. Horizon boundaries usually correspond to the boundaries of the formations, which commonly coincide with hiatuses.
Key words: **Acritarchs, biostratigraphy, Middle Cambrian, Upper Cambrian, East European platform.**

In the East European platform, Lower Cambrian rocks are the most widespread among the Cambrian deposits. The lower Cambrian strata crop out in the marginal parts of the shields, whereas in areas with monoclinal dipping of the western platform slope, they occur at somewhat deeper levels. The thickness of the Lower Cambrian deposits is up to 450 m and more.

The Middle and, in particular, Upper Cambrian rocks have a limited distribution in the region. These deposits are overlain with a thick sedimentary cover, although they also can be observed in natural outcrops located along the Baltic-Ladoga clint. The thickness of the Middle-Upper Cambrian strata is less than that of the Lower Cambrian. The Middle Cambrian is about 200 m thick, whereas the Upper Cambrian deposits are a few dozen of meters in thickness. The deep levels of these rocks make their study possible only by using core samples. This and poor paleontological records negatively influenced earlier investigations.

The age determinations for these deposits, which were considered in different areas to represent independent formations or beds (the Izhora, Sablino, Ladoga, Pestovo, and Vyzhev formations; the Obolidae, Kuvshinovo, Petseri, and other beds), ranged from the Early Cambrian to the Early Ordovician. No uniform correlation scheme existed either. The early investigations of these strata have been comprehensively discussed in several publications dedicated to this issue (e.g., Shestakova *et al.*, 1976; Plisov *et al.*, 1977). Still, only a few of them (Balashova, 1963; Khazanovich, 1968; and others) dealt with the paleontological substantiation of the Middle-Upper Cambrian subdivision in the Moscow syneclise and Baltic region. Intense detailed investigations of these strata were undertaken only in the 1980s in connection with oil and gas prospecting in the central and northwestern parts of the platform, namely, in the Moscow and southeastern Baltic syneclises (Kaliningrad region and western Lithuania). New finds of faunal remains reported during these works (Borovko *et al.*, 1980, 1984; Borovko and Sergeeva, 1981; Korobov and Yankauskas, 1982; Dmitrovskaya *et al.*, 1983) and acritarch studies (Volkova, 1980, 1982; Yankauskas, 1980; Volkova *et al.*, 1981; Kir'yanov *et al.*, 1991) provided reliable age substantiations for these strata. It was shown that they mostly correspond to the Middle-Upper Cambrian, and only the upper part of the Obolidae Beds correlated with the Tremadocian. Recently, the Cambrian-Ordovician boundary has been placed within the Obolidae Beds (Borovko *et al.*, 1984). Scarce faunal remains and limited core material have made acritarchs the principal fossil group in biostratigraphic studies on these deposits.

In this respect, publications by Volkova (1974, 1980, 1982, 1988, 1989a, 1989b, 1990a, 1990b), Volkova and Golub (1984, 1985), Yankauskas (1980, 1982, 1991), Bryzgalova (Nikashin *et al.*, 1986), Kir'yanov (Kir'yanov, 1993; Kir'yanov *et al.*, 1991; Kir'yanov and Prikhod'ko, 1994), and Paalits (1992) should be mentioned. These works provided a basis for elaborating the regional stratigraphic scheme discussed below. Studies that were performed by Vanguestaine (1978) and Marten and Dean (1981, 1984, 1988) in Belgium and Newfoundland, respectively, have contributed much to the elaboration of the scheme.

The regional scheme, which was presented at the Interdepartmental regional meeting on Cambrian deposits of the Russian platform, held in Vilnius in 1983, mainly included the Lower Cambrian horizons. In the Middle Cambrian, only two horizons, the Kibar-

tai and Deimena, representing its lower part, were distinguished at that time (*Reshenie...*, 1986).

The regional stratigraphic scheme of the Middle-Upper Cambrian deposits of the East European platform discussed in this paper (the figure) includes only those regions in which horizon stratotypes are established. To make the scheme more convenient, Cambrian sequences in Byelorussia and Poland are not represented here. Paleontological characteristics of the Middle and Upper Cambrian deposits in both of these countries are insufficient. The possible correlations of certain series and formations of Byelorussia and Poland are indicated in the text.

The suggested horizons are defined on the basis of acritarch assemblages, which were recorded in individual groups and formations or in their parts. These assemblages are not necessarily confined to the boundary layers of the formations; therefore, the layers incorporating acritarch assemblages were united with overlying and underlying strata barren of acritarchs. Sometimes, the horizon boundaries coincide with the geological boundaries of the formations, which are often limited by hiatuses. When a boundary between certain horizons is located within the formation, it is drawn in the middle between the acritarch assemblages that are correspondingly related to the two sequential horizons.

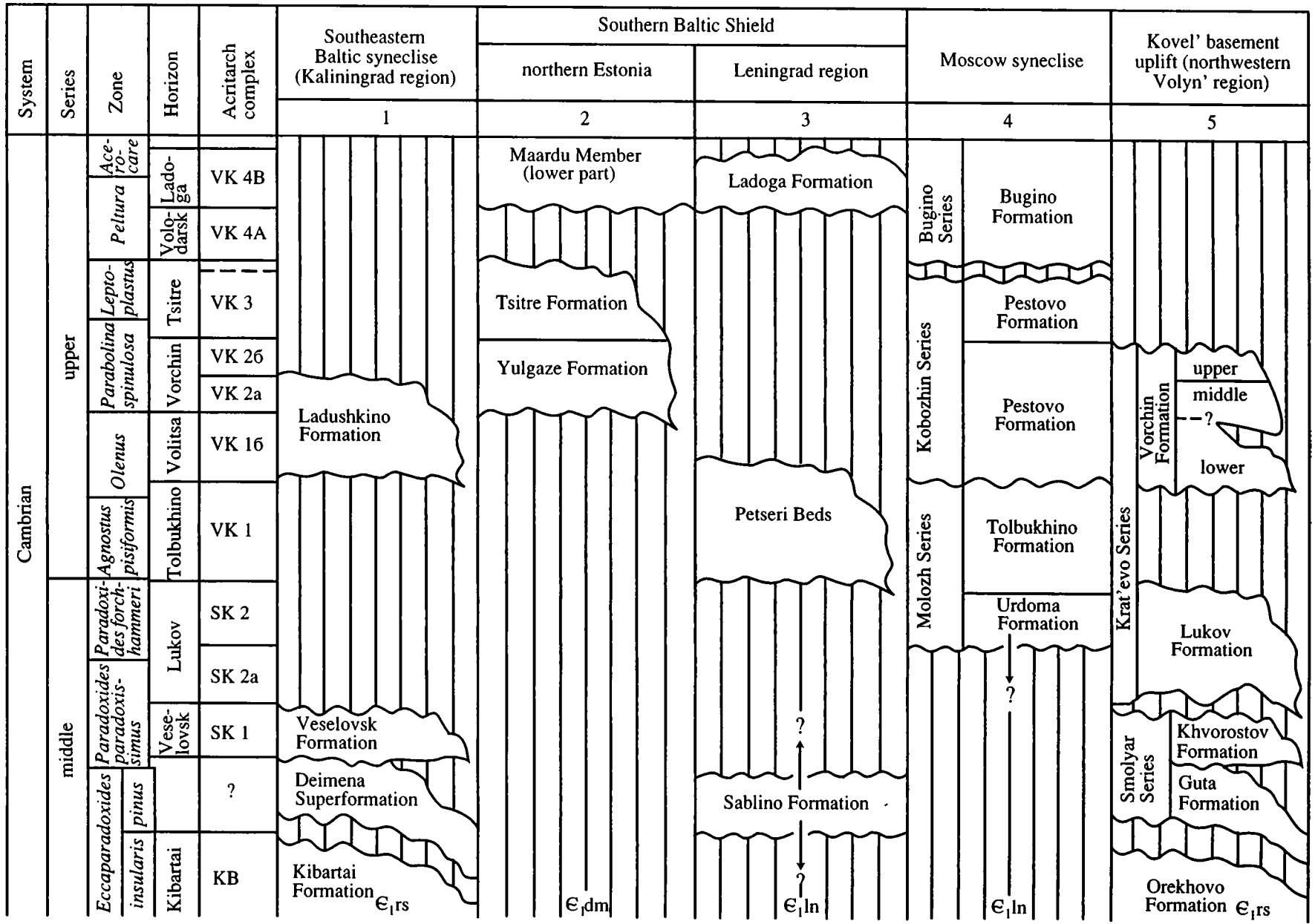
The Kibartai Horizon is the oldest Middle Cambrian horizon in the East European platform. Its name was introduced by Yankauskas (1974). A description of the horizon was given somewhat later (Brangulis *et al.*, 1975) when the paleontological characteristics were first identified. In subsequent years, they were merely specified. For instance, the acritarch forms, which were previously determined as *Cymatiosphaera* sp. 1 and *Cymatiosphaera* sp. 2, have recently been referred to *Cristallinum* sp. and *Eliasum llaniscum* Fombella, respectively, and they are considered to be the horizon reference species. Neither the stratotype sequence nor lateral facies changes were indicated at that time. Therefore, 1979 is adopted as the formal date of the registration of the horizon (Aren' *et al.*, 1979). The most complete information on this subject is presented in Kir'yanov (1991), Yankauskas (1982, 1991), and others.

The interval 1040.5 - 1077.8 in well Vergale-50, drilled in the southeastern Baltic syneclise, western Latvia, was adopted as the horizon stratotype. This interval corresponds to the Kibartai Member of the Tebra Formation. No paleontological description of the horizon stratotype has been published so far. Based on the studies of organic remains in coeval sediments penetrated by closely spaced wells (Vergele-49, Liepaya, and some others), the lower boundary of the horizon is defined by the appearance of acritarchs *Cristallinum* sp., *Baltisphaeridium pseudofaveolatum* Fridrichsone, *B. latviense* Volk., *Lophospaeridium variabile* Volk., and *Liepaina plana* Jank. et Volk. Also characteristic of the horizon are *Eliasum llaniscum* Fombella, *Micrhystridium notatum* Volk., and *M. obscurum* Volk. These latter are also present in the underlying deposits; however, their abundance in the Kibartai rocks sharply increases. In addition to the above-listed forms, there are species that, though present in other assemblages, are characteristic of the underlying horizons (the Vergele and Rausve horizons). These are abundant *Skiagia insigne* (Fridrichsone), *Micrhystidium lubomlense* Kirjan., *Volkovia dentifera* (Volk.), *V. conifera* (Jank.), *Aranidium confusum* Jank., and *Baltisphaeridium implicatum* Fridrichsone, which are associated with rare and single *Skiagia ciliosa* (Volk.), *Estiastra minima* Volk., *Leiovalia tenera* Kirjan., and some others (Volkova *et al.*, 1979).

The Kibartai deposits also contain trilobites *Ellipsocephalus politomus* (Linnarson), *E.* cf. *puschi* Orlowsky, *Strenuaella* (*Comluella*) aff. *samsonowichi* Orlowsky, and *S.* (Comluella) *insolita* N. Tchern. (Lendzen *et al.*, 1973; Mens *et al.*, 1987) as well as small obolids.

The Kibartai Horizon deposits conformably overlie the older layers. As was mentioned, the lower boundary of the horizon is biostratigraphically distinct and defined by the replacement of the Rausve acritarch assemblage with the Kibartai one. The terminal Kibartai time was evidently marked by a structural platform reorganization, and, as a result, the Kibartai deposits were overlain with an azimuthal unconformity by younger Cambrian strata (Brangulis *et al.*, 1975; *Reshenie...*, 1986; Kir'yanov *et al.*, 1991; Kir'yanov and Prikhod'ko, 1994). The depositional environments also changed sharply (evidently, they became more shallow-water) at that time. Meanwhile, the upper boundary of the horizon bears no distinct biostratigraphic features to make its definition unambiguous, because, in this case, we can hardly apply the principle according to which the upper boundary of the underlying unit is defined by the level of the overlying unit lower boundary. Thus, in most sequences, this boundary is recorded according to the replacement of the diverse (in respect to both genera and species) Kibartai acritarch assemblage with microphyte fossils consisting of numerous phytoleim fragments and rare thick-walled leiosphaerids. However, there are some sequences in which lithological changes are associated with the impoverishment of the Kibartai acritarch assemblage composition, but not with the aforementioned changes (Kir'yanov *et al.*, 1991; Yankauskas, 1991). Most likely, this phenomenon can be explained by the diachronous character of the lower boundary of the post-Kibertai facies (the figure). Therefore, there is no point in bounding the Kibartai Horizon from the top by the level of the "Kibartai" facies (Yankauskas, 1991), because such an approach would mask the "slipping age" of the considered boundary.

Throughout the platform, this horizon corresponds to the following units: the Kibartai Formation in western Lithuania and the Kaliningrad region; the upper (Kibartai) member of the Tebra Formation in western Latvia and the southeastern Baltic syneclise; the Stavsk



Regional stratigraphic scheme of Middle-Upper Cambrian deposits in the East European platform.
Underlying Lower Cambrian deposits in some areas are indicated by letter indices: $\epsilon_1$ln – Lontova deposits, $\epsilon_1$dm – Dominopole deposits, $\epsilon_1$rs – Rausve deposits.

Formation in the eastern Podlyassk-Brest depression of Byelorussia; the Orekhovo Formation and (sometimes) lower Guta Formation in the Kovel' basement uplift in the northern Lviv depression of Ukraine; and, probably, the upper Radzyn' Formation in the Lyublin slope area of the platform and in the western Podlyassk-Brest depression of eastern Poland.

In age, the Kibartai Horizon is correlative with the Middle Cambrian trilobite *Eccaparadoxides insularis* Zone of the West European zonal scale (Mens *et al.*, 1987).

In some areas of the East European platform, the Kibartai deposits are overlain by a thick (more than 60 m) member, which is mainly composed of sandstones nearly barren of fossils. This member is represented by the following: the Deimena Group (Superformation) in western Lithuania, Latvia, and the Kaliningrad region; probably the Rukhnu Beds of western Estonia; and the middle Lyna Formation in the southeastern Baltic syneclise, eastern Poland. Similar rocks also compose the following the Orlya Formation in the eastern Podlyassk-Brest depression of Byelorussia; the lower Kostzhin' Formation (probably the entire unit) in the western Podlyassk-Brest depression and in the Lyublin slope area of the platform in eastern Poland; and, finally, the Guta Formation in the Kovel' basement uplift and the northern Lviv depression of western Ukraine.

As was mentioned, some of the above-listed units were considered to represent the single Deimena Horizon (*Reshenie...*, 1986; Yankauskas, 1991). However, in our opinion, there are no sufficient grounds thus far for uniting all these members into a single regional horizon. Therefore, in the proposed scheme, we leave a gap corresponding to the interval between the Kibartai Horizon and the Middle Cambrian Veselovsk regional horizon, which is proposed here for the first time. If further investigations confirm that the stratigraphic interval corresponding to the gap in the scheme represents a single regional horizon, then it could retain the name "Deimena Horizon."

The deposits that directly overlie in some platform areas (the figure) and the above-considered fossil-free Middle Cambrian strata are considered here as the Veselovsk Horizon. The stratotype of the Veselovsk Formation, interval 2463.6 - 2468.4 m in the Veselovsk-8 well drilled near the Veselovsk Settlement in the Kaliningrad region, was adopted as the stratotype for this horizon.

The lower boundary of the horizon is defined by the appearance of acritarchs *Adara* sp. and larger forms of *Eliasum* sp. 1. The Veselovsk acritarch assemblage, which is also known in the literature as the acritarch complex SK 1 (Volkova, 1990a), is characterized by *Cristallinium cambriense* (Slavikova), *Micrhystridium* aff. *lanceolatum* Vang., *M.* spp., *Celtiberium*? sp., *Multiplicisphaeridium* sp., *Ovulum* sp., *Betisphaeridium* sp., and *Zonosphaeridium* cf. *ovillensis* Cramer et Diez, associated with smaller *Tasmanites* sp. and thick-walled leiosphaerids as well.[1]

The horizon strata rest with a hiatus upon the older Middle Cambrian deposits and, in turn, are unconformably overlain by younger layers that vary in age, including ones from the Lower Ordovician (the figure).

The correlatives of this horizon are the Veselovsk Formation in the Kaliningrad region of the southeastern Baltic syneclise (Nikashin *et al.*, 1986) and the Khvorostov Formation of the Kovel' basement uplift in western Ukraine (Kir'yanov and Prikhod'ko, 1994), where *Eliasum* sp. 1, *Cristallinium cambriense*, and *Adara* sp. occur, which are typical forms of this horizon.

Acritarch assemblages similar in composition to those recorded in our samples (microfloras A0 and A1) were also reported from the middle, thickest part of the Manuels River Formation in eastern Newfoundland (Martin and Dean, 1984, 1988). Based on the trilobite assemblage, these deposits are referred to the middle *Paradoxides paradoxissimus* Superzone of the West European scale (excluding the *Ptychagnostus gibbus*, the lowermost *P. atavus*, and probably the *P. punctuosus* zones). The age of the deposits with the Veselovsk acritarch assemblage can be specified in the course of further studies.

In the most complete Middle Cambrian sequences of the Volyn' region, the Veselovsk Horizon deposits are overlain by strata that we consider as the Lukov Horizon. The interval 380.7 - 385.5 m in well 5653, drilled in the village of Staraya Guta, the Volyn' region, is considered to represent a stratotype of this horizon; the interval 325.2 - 333.1 m in well 5457 in the village of Lukov (stratotype of the Lukov Formation) was chosen as its parastratotype.

The lower boundary of the horizon is marked by the appearance of the acritarch assemblage with characteristic species *Timofeeevia janischewskyi* (Timofeev), *T. ancarae* (Cramer et Diez), *Cristallinium dubium* Volk., and *Dictyotidium* sp. (aff. *D. hasletianum*). *C. cambriense*, *Solisphaeridium flexiplosum*, *Comasphaeridium* sp., and some other acritarch forms. In the Moscow syneclise, this acritarch association is known as the complex SK 2 (Volkova, 1990a). The acritarch assemblage recorded in the western platform sequences (Kovel' basement uplift) is subdivided into two subassemblages. The lower one, which is designated as SK 2a in the proposed scheme, contains abundant (often dominating) small acritarch forms:

[1] In addition to the listed acritarch genera and species, the representatives of the *Timofeevia* genus were reported from the Veselovsk Formation, specifically from its stratotype (Nikashin *et al.*, 1986; Yankauskas, 1991). These data contradict our materials (including the samples collected from the stratotype), that show that the considered deposits lack species of the *Timofeevia* genus. According to our observations, they appear in younger Middle Cambrian strata. Taking into account the fact that no illustrations of the *Timofeevia* forms were given in the cited publications, we eliminated these data from our analysis.

Micrhystridium spp. (including, probably, new species), *M. lanceolatum* Vang., *M. lubomlense* Kirjan., *Lophosphaeridium*? *kryptoradiatum* Vang., *Ovulum* spp., *O. saccatum, O. lanceolatum* Jank., *Volkovia* sp. (aff. *V. dentifera, V. flagellata, Aranidium ihoricum* Jank., and *A. aculeatum* Jank. (Kir'yanov, 1983).

In addition to acritarchs, the sediments of the Lukov Horizon show the presence of trilobite *Agnostus subsulcatus* West. (Korobov and Yankauskas, 1982). In the Volyn' region, abundant large-shelled Obolidae occur in these rocks.

The Lukov Horizon sediments transgressively cover the underlying strata and are conformably, or with a hiatus, overlain by Upper Cambrian strata that vary in age (the figure). The horizon comprises sediments of the Urdoma and lower Tolbukhino formations in the central part of the Moscow syneclise, the Lukov Formation of the Kovel' basement uplift in western Ukraine, the Genchai Member of the southeastern Baltic syneclise in western Lithuania, and the Buikovo Formation (probably its upper part only) in the northeastern slope of the Byelorussia-Mazur anticlise. Some investigators also refer to this horizon as the Paneryai Formation, which is distinguished in eastern Lithuania (Yankauskas, 1991). However, the occurrence of some species of the Paneryai acritarch assemblage, such as *Micrhystridium lanatum* Volk., *M. dissimilare* Volk., *Skiagia* aff. *ciliosa* (Volk.). *Alliumella baltica* Vanderflit, and *Comasphaeridium strigosum* (Jank.), which were not recorded elsewhere above the Kibertai strata, contradicts this conclusion. This fact also prevents us from using the term "Paneryai Horizon" for the considered deposits.

The difference between the two acritarch subcomplexes from the Lukov Horizon in the Kovel' basement uplift led to the conclusion that the Lukov Formation is not the complete everywhere; some sequences lack the lower part (Kir'yanov and Prikhod'ko, 1994).

The considered deposits correlate with the trilobite *Paradoxides forchhammeri* Superzone (uppermost part excluded) and, probably, to the upper *P. paradoxissimus* Superzone (Volkova, 1990a).

We named the next, overlying unit the Tolbukhino Horizon. The stratotype is found in well Tolbukhino-1 (interval 2149 - 2173.4 m), drilled in the vicinity of Tolbukhino Village, 10 km west of Yaroslavl'.

The lower boundary of the horizon is defined by the appearance of acritarchs *Raphesphaera turbata* (Martin), *R. spinulifera* Volk., *Timofeevia pentagonalis* Vang., and *Pirea orbicularis* Volk. and *Poicilofusa* sp. *Timofeevia janischewskyi, T. lancarae, Cristallinium cambriense, Dictyotidium* sp., and *D.* aff. *hasletianum*. Some other species characteristic of the underlying deposits continue to occur in this horizon. A similar assemblage was recorded in the Tolbukhino Formation of the Moscow syneclise, where it is known as the complex VK 1 (Volkova, 1990a). In addition to acritarchs, the Tolbukhino sediments yielded the remains of inarticulate brachiopods belonging to the *Ungula, Oepikites*, and *Angulotreta* genera.

The horizon deposits rest, conformably or with a hiatus, upon the underlying Lower and Middle Cambrian strata and are overlain by the younger Upper Cambrian rocks. The boundary with the latter has a conformable or erosional character. The Tolbukhino Horizon is correlative with the Petseri Formation with the acritarch complex VK 1 (Volkova, 1990a) in the southern Baltic shield (the eastern Leningrad region) and with the largest upper part of the Tolbukhino Formation in the central area of the Moscow syneclise. In the well Tolbukhino, the acritarch-free interval between the lower and upper parts of the Tolbukhino Formation (i.e., between the Lukov and Tolbukhino regional horizons) is about 5 m.

The horizon deposits are correlative in age with the upper part of the trilobite *Paradoxides forchhammeri* Superzone, the *Agnostus pisiformis* Zone, and approximately with the lower half of the *Olenus* Zone (Volkova, 1990a).

The overlying Cambrian strata of the East European platform are identified as the regional Volitsa Horizon, with the stratotype corresponding to the interval 249.0 - 252.3 m in well 5743 drilled in the village of Volitsa of the Volyn' region, western Ukraine.

The appearance of acritarch species *Cymatiogalea* sp. (*C.* aff. *aspergillum*) and *C. virgulta* Martin marks the lower boundary of this horizon. The acritarch assemblage also includes *Timofeevia* sp., *T. janischewskyi, T. lancarae, Retisphaeridium dihamerum* Staplin, Jansonius et Pocock, *Raphesphaera spinulifera, Cristallinium cambriense*, and *Synsphaeridium* sp. In our scheme, this assemblage is designated as the complex VK 1b (the figure).

The horizon deposits unconformably overlie the older strata and, in turn, are overlain (probably with a hiatus) by younger Upper Cambrian and, in some sequences, Lower Ordovician members.

The horizon comprises the lower Vorchin Formation of the Volyn' basement uplift in western Ukraine and, evidently, the lower Ladushkino Formation of the southeastern Baltic syneclise in the Kaliningrad region.

The appearance of forms with a large polar foramen (*Cymatiogalea*) is a distinctive feature of the Volitsa acritarch assemblage. A similar assemblage with representatives of the *Cymatiogalea* and *Stellioferidium* genera was reported to occur in the Cambrian sequences of Great Britain and Newfoundland beginning from the basal (Potter, 1974) or middle part (Martin and Dean, 1988) of the *Olenus* Zone, respectively. This suggests that the lower age limit of the deposits with the Volitsa acritarch assemblage can not be older than the middle level of this zone. At the same time, the Volitsa assemblage lacks *Veryhachium dumontii* Vang. and *Leiofusa stoumonensis* Vang., which appear in eastern Newfoundland in the upper part of the trilobite-free member between the layers that are referred there to the *Olenus*

and *Parabolia spinolosa* zones, respectively. Proceeding from this evidence, we correlate the upper boundary of the horizon with the boundary layers between these two zones, and we correlate the horizon itself with the upper portion of the *Olenus* Zone.

A similar age can evidently be assigned to the lower Ladushkino Formation, where trilobite *Homagnostus* and brachiopod *Orusia lenticularis* were recorded (Nikashin *et al.*, 1986). Based on this evidence, the lower part of the above formation is included in the Volitsa Horizon.

Higher Upper Cambrian strata in the East European platform compose the Vorchin regional horizon. The Vorchin Formation stratotype, which was penetrated in the interval 426.5 - 437.6 m of well 5405, drilled in the village of Kret' in the Volyn' region of western Ukraine (Kir'yanov and Prikhod'ko, 1994), was adopted to represent the horizon stratotype.

The lower boundary of the horizon is defined on the basis of the appearance of acritarchs *Leiofusa stoumonensis*, *Verihachium dumontii*, *Cristallinium baculatum* Volk., *Cymateogalea fimbriata* Volk., *C. cristata* (Downie), *C. velifera* (Downie), *Raphosphaera obsoleta* Volk., *Stelliferidium cortinulum* (Deunff), *Timofeevia estonica*, and *Impluviculus multiangularis* (N. Umn.) (Volkova, 1990a). *Timofeevia janischewskyi*, *T. lancarae*, *Raphoshaera turbata*, *R. spinulifera*, *Cymbatiogalea* sp. (*C.* aff. *aspergillum*), *C. virgulta* and other acritarch species characteristic of underlying deposits continue to occur in this horizon. In the upper third of the horizon, *Dasydiocrodium caudatum* Vang. and *Stellochinatum uncinatum* (Downie) appear, which provided grounds for subdividing the Volitsa acritarch assemblage (VK 2, according to Volkova) into two subassemblages: the lower Vorchin (VK 2a) and the upper Vorchin (VK 2b). In addition to acritarchs, the horizon sediments in the western East European platform yielded abundant shells of inarticulate brachiopods *Ungula inornata* (Mickwitz), *Oepikites fragilis* Popov et Khazanovich, *Angulotreta postapicalis* Palmer, and *Ceratreta tanneri* (Metzger) and conodonts *Furnishina alata* Szaniawski, *F. furnishi* Muller, *Proonectodus tenuis* (Müller) and *Westergaardodina bicuspidata* Muller (*Opornye...*, 1989).

The Vorchin Horizon deposits rest, usually unconformably, upon the older strata and are overlain, conformably or with a hiatus, by the Upper Cambrian and Lower Ordovician rocks, respectively. The horizon comprises the middle and upper subformations of the Vorchin Formation in the Kovel' basement uplift, western Ukraine, and the Yulgaze Formation and upper Petseri Formation in the southern Baltic shield, northern Estonia. The sediment member, which in eastern Poland is referred to the Tremadocian [e.g., the interval 4423.7 - 4460.0 m of well Lopiennik IG-1 with *Leiofusa* sp. (*L.* aff. *stoumonensis*), *Stelliferidium cortinulum*, *C. virgulta*, and other acritarchs (Kir'yanov and Prikhod'ko, 1994)], should undoubtedly be included (partly or entirely) in the horizon. The upper Ladushkino Formation of the southeastern Baltic syneclise in the Kaliningrad region probably also corresponds to this regional unit (Nikashin *et al.*, 1986)

In age, the considered deposits approximately correspond to the lower trilobite *Parabolina spinulosa* Zone (Volkova, 1990a, 1990b).

The next Upper Cambrian horizon was named after the Tsitre Formation, which occurs in northern Estonia (*Opornye...*, 1989). The horizon stratotype corresponds to the interval 112.8 - 119.6 m of well M-72, drilled 30 km southeast of Tallinn (Faalits, 1992).

The lower boundary of the horizon is marked by the appearance of acritarchs *Acanthodiacrodium* spp. and *Trunculamarium revunum* (Vang.); species of the *Lusatia* genus; and abundant, though so far poorly studied representatives of the *Polygonum* (=*Goniosphaeridium*) genus. Species *Veryhachium dumontii*, *Leiofusa stoumonensis*, *Dasyodiacrodium caudatum*, and *Raphosphaera turbata* as well as forms of *Timofeevia*, *Cristallinium*, *Cymatiogalea*, and other genera pass into this horizon from older deposits. The Tsitre acritarch assemblage is known in the literature as the complex VK-3 (Volkova, 1990a). In our scheme, it has the same designation. Obolids *Oepikites obtusus* (Mickwitz), *Schmidtites celatus* (Volborth), and *Ungula ingrica* (Eichw.) and conodonts *Phakelodus tenuis* (Müller), *Procnoctudus* cf. *galeatini* (Müller), *Westergaardodina bicuspidata*, and others supplement the paleontological characteristics of the horizon (*Opornye...*, 1989).

The Tsitre Horizon deposits transgressively cover either the Vorchin Horizon or the Lower Cambrian layers, and are overlain with a hiatus by the Upper Cambrian strata. At present, the horizon includes only the Tsitre Formation of the southern Baltic shield, northern Estonia. Previously, the latter was considered to correspond to the upper Yulgeze Formation of certain sections (*Opornye...*, 1989). Its spatial distribution is confined to the area south of the Maardu deposit. In age, the Tsitre Horizon corresponds to the upper *Parabolina spinulosa* and *Leptoplastus* trilobite zones (Volkova, 1990a; Paalits, 1992).

The interval 1840.0 - 1856.0 m in well Rybinsk-1, drilled in the vicinity of Rybinsk in the Yaroslavl' region, is adopted to represent a stratotype of the next Upper Cambrian unit, the Volodarsk Horizon.

The appearance of acritarchs *Impluviculus villosiusculus* Volk., *Lusatia* sp. 1, *L. triangularis* (N. Umn.), and *Striatitheca loculifera* Volk. ascertains the position of the lower boundary of the horizon. Also typical of these deposits are *Dasydiacrodium caudatum*, *Leiofusa stoumonensis*, *Veryhachium dumontii*, and *Trunculamarium revinum*, as well as representatives of *Timofeevia*, *Raphosphaera*, *Cymatiogalea*, and *Acanthodiacrodium* genera and other acritarch species, which occur in older strata. The Volodarsk acritarch assemblage was previously described as the complex VK 4A

(Volkova, 1990a). Associated trilobite remains (*Parabolina* sp.) are typical, in Lendzen's opinion, of the *Peltura* Zone. The lower Bugino Formation in the central Moscow syneclise (recovered in the wells Rybinsk-1 and Danilovo-11), where it unconformably overlies the older strata, belongs to this horizon.

As was mentioned, the find of trilobite *Parabolina* sp. suggests that the Volodarsk Horizon is correlative with the *Peltura* Zone. In the central Moscow syneclise, however, the latter also includes the higher Upper Cambrian strata, which contain a quite different acritarch assemblage (VK 4B). Therefore, we tentatively refer the interval with the Volodarsk acritarch assemblage (VK 4A) to the lower part of the *Peltura* Zone (the group of zones in a broad sense).

The uppermost Middle-Upper Cambrian regional unit that is suggested here was named the Ladoga Horizon, with its stratotype identical to that of the Ladoga Formation (Borovko *et al.*, 1984).

The lower boundary of the horizon corresponds to the appearance level of acritarchs *Izhoria angulata* Golub et Volk., *Ocidium rossicum* Tim., *Dasydiacrodium palmatilobum* Tim., *Arbusculidium destombesii* Deunff, *Calyxiella izhoriensis* Golub et Volk., *Nellia longiuscula* Golub et Volk., *N. magna* Volk., *Vogtlandia notabilis* Volk., *Schizodiacrodium brevicrinitum* Golub et Volk., *S. fibrosum* Golub et Volk., and other species. The Ladoga assemblage (VK 4B) also includes species of *Cymatiogalea*, *Stelliferidium*, and *Timofeevia* and other acritarch genera typical of older horizons (Volkova, 1990a). In addition, trilobite *Parabolina jaroslavica* Suvorova (Shestakova *et al.*, 1976) and a diverse assemblage of inarticulate brachiopods and conodonts (*Opornye...*, 1989) were reported from these deposits.

In both the stratotype locality and the southern Baltic shield, northern Estonia, the Ladoga Horizon has an erosional contact with the Middle-Upper Cambrian strata and, with a hiatus, is overlain by the Lower Ordovician sediments. In the Moscow syneclise, the stratigraphic relationship of the horizon with both the underlying and overlying strata is uncertain because of poor recovery. The horizon incorporates the Ladoga Formation and the lower Maardu Member in the southern Baltic shield (Leningrad region and northern Estonia) as well as the middle part of the Bugino Formation in the Moscow syneclise.

Based on the find of trilobite *Parabolina jaroslavica*, which is most likely confined to the upper *Peltura* s.l. Zone (or *Peltura scarabicides* Zone), as well as an acritarch assemblage (VK 4B) identical to that from the *Peltura scarabocides* Zone of Sweden (Tongiorgi and Ribecai, 1990), the Ladoga Horizon should be referred to the upper *Peltura* Zone.

In the northwestern Russian platform, the Ladoga Horizon deposits are overlain, conformably or with a hiatus, by the lower Tremadocian. Its upper boundary is defined by the appearance of acritarch species characteristic of the lower Tremadocian: *Acanthodiacrodium angustum* (Downie) Combas, *A. comptulum* Rasul., *Dicrodiacrodium ramusculosum* (Combas) Volk., and others. The appearance level of the listed species and, accordingly, the Ladoga Horizon upper boundary approximately coincide with the base of the conodont *Cordylodus proavus* Zone. The sediments of older *Proconodontus* and *Cordylodus andresi* conodont zones yielded an acritarch assemblage similar to that from the Ladoga Horizon (Volkova and Mens, 1988; Volkova, 1989b). Typical Upper Cambrian acritarch species *Calyxiella izhoriensis*, *Ocidium rossicum*, *Vogtlandia notabilis*, and others become extinct near the Ladoga Horizon upper boundary.

Thus, based on our data, the Middle Cambrian of the East European platform can be subdivided into the Kibartai, Veselovsk, and Lukov regional stratigraphic horizons. The Tolbukhino, Volitsa, Vorchin, Tsitre, Volodarsk, and Ladoga regional horizons are defined in the Upper Cambrian deposits of the platform. Except for the Kibartai Horizon, all of the listed Middle-Upper Cambrian regional stratigraphic horizons are established here for the first time.

The subdivision into horizons of the Phanerozoic sedimentary sequences, which were formed in a single paleobasin, was generally accepted by geological surveys of different republics of the former USSR. We hope that the regional scheme proposed in this article for the subdivision of Middle-Upper Cambrian sequences will be applicable in geological studies that are being carried out in the territory of East European platform by geologists of countries that are now independent.

Reviewer A.Yu. Rozanov

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Paleozoic Deposits in Kolguev Island (Barents Sea)

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Abstract – Based on the results of a study of Paleozoic deposits penetrated by boreholes in Kolguev Island, a paleontologically substantiated stratigraphic scheme is proposed for their subdivision for the first time. Stages and some substages of the general scale are established. This sequence can be used as a reference for seismic works and drilling in the adjacent shelf areas, where oil and gas deposits have been discovered.

Key words: **Barents Sea, Kolguev Island, boreholes, stratigraphy, lithology, Paleozoic, correlation, facies.**

The Barents Sea recently attracted great attention because of the discovery of the Prirazlomnoe oil field, the huge Shtokman, and a series of smaller gas deposits there. Kolguev Island is a single, rather large (80 × 80 km) land area located in the southwestern Barents Sea (Fig. 1A), and its structure is of special interest. It is almost entirely covered by Quaternary sediments. Upper Cretaceous rocks crop out only along the steep northern coast.

Only two publications contain some information on the Paleozoic deposits of the island. The first one (Entsova *et al.*, 1981) deals with the Upper Permian sequence (beginning with the Artinskian Stage), which is penetrated by two prospecting boreholes in the vicinity of the Bugrino Settlement on the southern coast of the island. The second paper (Bro *et al.*, 1988) gives a brief description, without paleontological characteristics, of the entire Paleozoic sequence, including the Ordovician, which was drilled by two boreholes in the Peschanoe Ozero structure in the eastern part of the island. There are also two small articles dealing with the Permian reefs (Preobrazhenskaya *et al.*, 1991) and Devonian palynofossils (Dibner *et al.*, 1988).

The discovery of oil in the Triassic deposits of the Peschanoe Ozero field promoted extensive deep drilling there, which was initiated in 1980. More than 50 boreholes with a maximum penetration of 4740 m (Borehole WP-1) have been drilled there since that time.

The boreholes are unevenly spaced and mainly confined to the Peschanoe Ozero field in the eastern part of the island (Fig. 1B); drilling was also performed at seven other sites. Almost all boreholes recovered the Upper Paleozoic strata, several of them penetrated the Devonian deposits, and two boreholes were terminated in the basal Ordovician layers (BG-1 in the south and NW-202 in the northwest of the island).

Recovery varied from 7.5 to 18%. The closely spaced distribution of holes in the Peschanoe Ozero oil field allowed the whole section to be sampled. High-quality logging permitted the correlation of all the drilled sequences to be made. The validity of the correlations was always tested by core analysis; fortunately, this was successful, because core recovery was best in those fields in which boreholes were rare.

Figure 2 shows the integrated Paleozoic sequence for the corresponding fields in Kolguev Island. The most complete Upper Permian section is illustrated by the more detailed columns of three boreholes in the southern, northwestern, and eastern parts of the island, respectively (Fig. 3).

A lithological analysis and compilation of sequences using logging data were performed by Preobrazhenskaya and Bro. Paleontological materials were analyzed by a large team of paleontologists from different institutions (All-Russian Research Institute "Okeangeologiya"; All-Russian Research Institute of Prospecting Geology; All-Russian Geological Institute; Paleontological Institute, Russian Academy of Sciences; and others). Determinations of different fossil groups were made by the following specialists: Rudakovskaya (Ordovician acritarchs); Popov and Bondarev (Ordovician brachiopods); Cherkesova (Devonian brachiopods); Petrosyan (Devonian floral remains); Ivanov (Devonian fishes); Dibner, Stanichnikova, Vasil'eva, and Verbova (Devonian miospores); Deulin and Sobolev (Devonian and Carboniferous conodonts); Solov'eva, Izotova, Nikolaev, Davydov, and Chuvashov (Devonian, Carboniferous, and basal Permian fusulinids); Sosipatrova and Grinchenko (Carboniferous and Permian small foraminifers); Chernyak and Ustritskii (Carboniferous and Permian brachiopods); Morozova and Nekhorosheva (Permian briozoans); Lobanova, Astaf'eva, and Kanev (Permian bivalves); Bro (Permian floral remains); Fefilova and Varyukhina (Permian pollen and spores); and Molin (Devonian and Permian phyllopods).

These studies allowed a detailed subdivision of the sedimentary cover to be carried out. Stages were established in all systems, and substages were defined for the

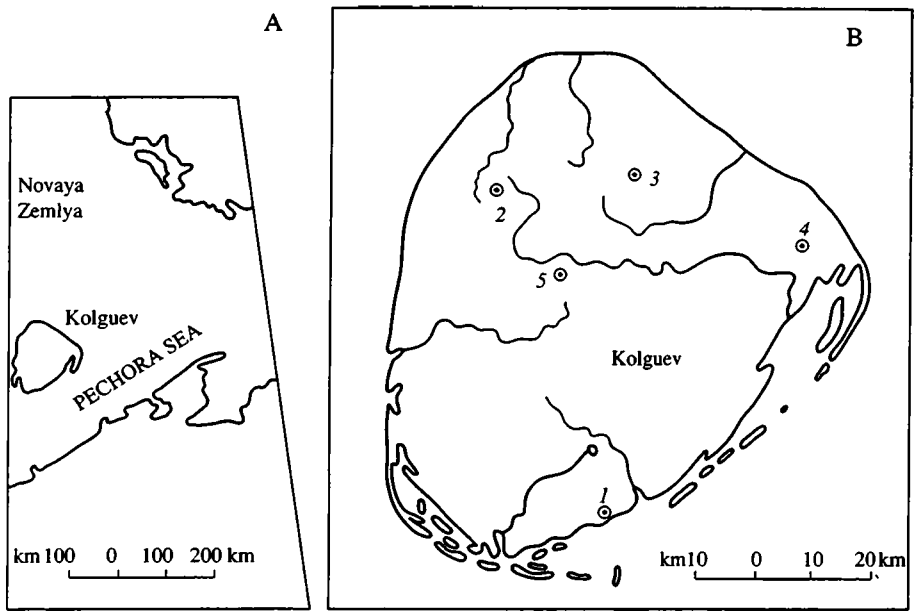

Fig. 1. A. Location of Kolguev Island in the Pechora Sea (southeastern Barents Sea).

B. Location of drilled sites in Kolguev Island.

Areas: *1* – Southern (boreholes Bugrino-1, 140, and 141), *2* – Northwestern (boreholes Northwestern-202 and Portnovo-1), *3* – Northern (borehole western Peschanoe Ozero-1), *4* – Eastern (boreholes Peschanoe Ozero 1 to 46), *5* – Central (boreholes Tarka 1 to 7).

Abbreviated borehole designations in the text:

Bugrino – BG, Northwestern – NW, western Peschanoe Ozero – WP, Peschanoe Ozero – PO, Tarka – T.

Devonian through the Carboniferous. The tentative stage definition for the Upper Permian is related to the imperfection of the Permian stage scale rather than to insufficient paleontological data. Good recovery in some boreholes made it possible for the horizons of the general scheme to be outlined. Nevertheless, correlation between all holes was possible only at the stage level. This, in turn, allowed some significant facial changes to be revealed, despite the small dimensions of the island, and it also allowed their nature to be explained. The results of these studies formed the basis of this article.

The Paleozoic sequence of Kolguev Island is distinctly divisible into six lithologic-stratigraphic complexes: (1) Ordovician, (2) Lower Devonian, (3) Upper Devonian, (4) Lower Carboniferous, (5) Middle Carboniferous-Sakmarian, and (6) Artinskian-Tatarian (Fig. 2). The four lower complexes are clearly separated from each other by the stratigraphic hiatuses between them. Despite its conformable contact with the underlying deposits, the uppermost complex is also easily recognized owing to its predominantly terrigenous composition. Nearly all of the deposits are horizontally layered.

1. The Ordovician complex is most completely represented in the southern and northwestern parts of the island. In the Peschanoe Ozero field, some boreholes (PO-4 and PO-46) penetrated only its uppermost layers. The complex comprises a thick terrigenous member with a stratigraphic range from the Cambrian-Ordovician boundary to the Middle(?) Ordovician. Its section represents a transgressive-regressive cycle, the lower part of which (Tremadocian Stage) is mainly composed of gray lagoon-marine deposits with acritarchs, brachiopods, and crustaceans (phyllocarids). The middle portion of the complex (Arenigian Stage) consists of alternating red, green, and sometimes grey marine and continental claystones, siltstones, and sandstones with acritarchs and brachiopods; rare trilobites occur in the basal Arenigian layers. The upper part of the cycle [Middle(?) Ordovician] is mostly made up of continental deposits. They are represented by a cyclic alternation of red and, less commonly, green siltstones, sandstones, and sometimes claystones.

The Tremadocian portion of the complex shows a cycle of the second order, the largest part of which is composed of dark-grey fine-grained siltstones that are replaced by pink sandstones and siltstones in the upper portion of the cycle. *B. exinatum*, *Tetraniveum* sp., *Dasydorus* sp., and *Peteinosphaeridium* sp. (Borehole BG-1, depth interval 3859 - 3783 m) and brachiopod species *Angarella* ex gr. *lopatini* Assat. (Borehole BG-1, depth 3781 m) are recorded. The thickness is 220 m.

The Middle(?) Ordovician portion of the complex represents the upper, regressive half of the large Lower-Middle Ordovician cycle of the first order. These layers are barren of fossils. Lithologically, they are mostly composed of red-colored siltstones and sandstones

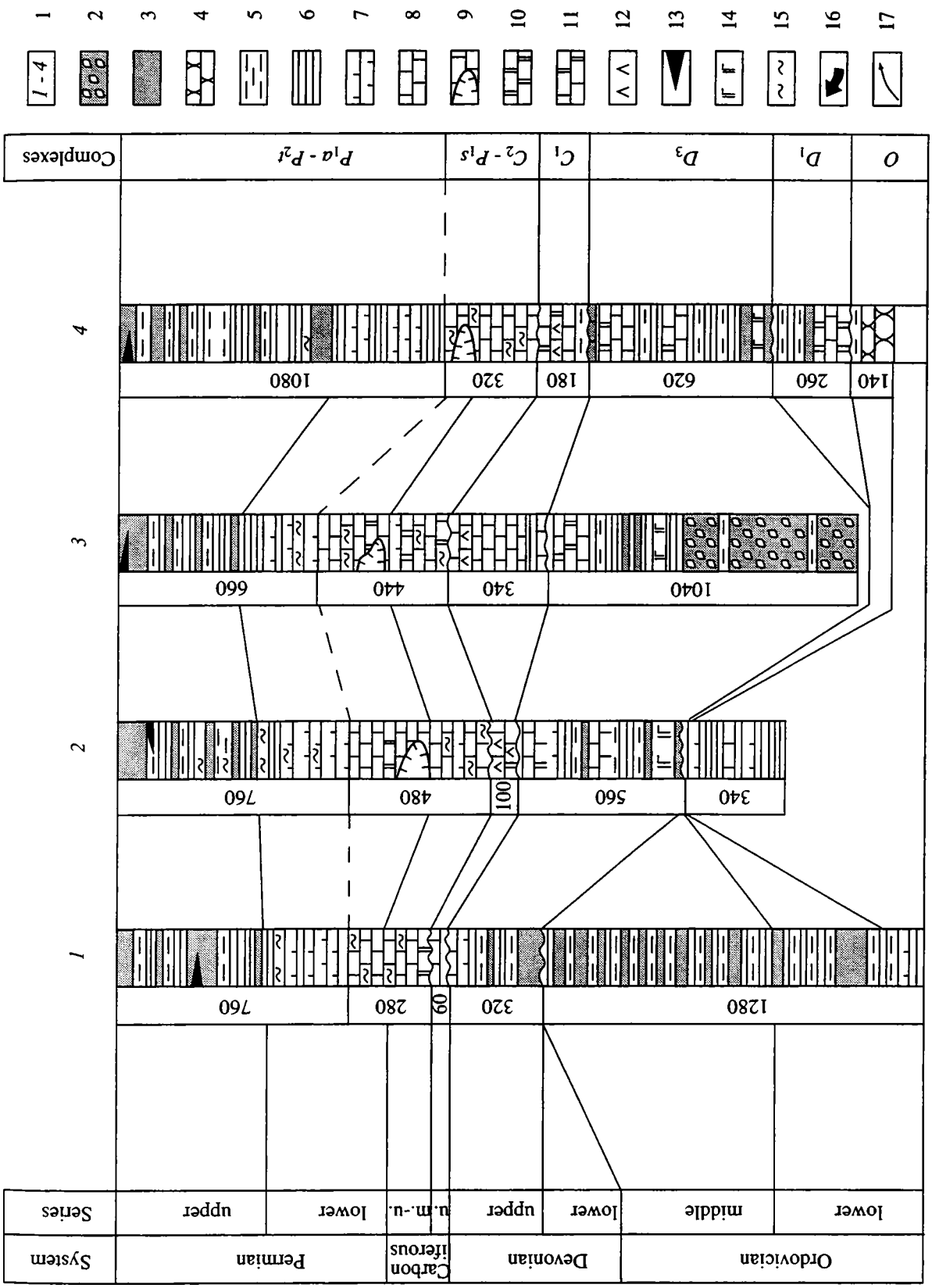

Fig. 2. Correlation of Paleozoic sequences (section numbers correspond to site numbers in Fig. 1).
1 – Sections, 2 – conglomerate, gravelstone, 3 – sandstone, 4 – quartzitelike sandstone, 5 – siltstone, 6 – claystone, 7 – calcareous claystone, marl, 8 – limestone, 9 – reefs and their clastic products, 10 – dolomite, 11 – dolomitized limestone, 12 – anhydrite, 13 – coal, coaly rocks, 14 – basalt, 15 – silicified rocks, 16 – transgressive part of the cycle, 17 – regressive part of the cycle (16 and 17 refer to Fig. 3).

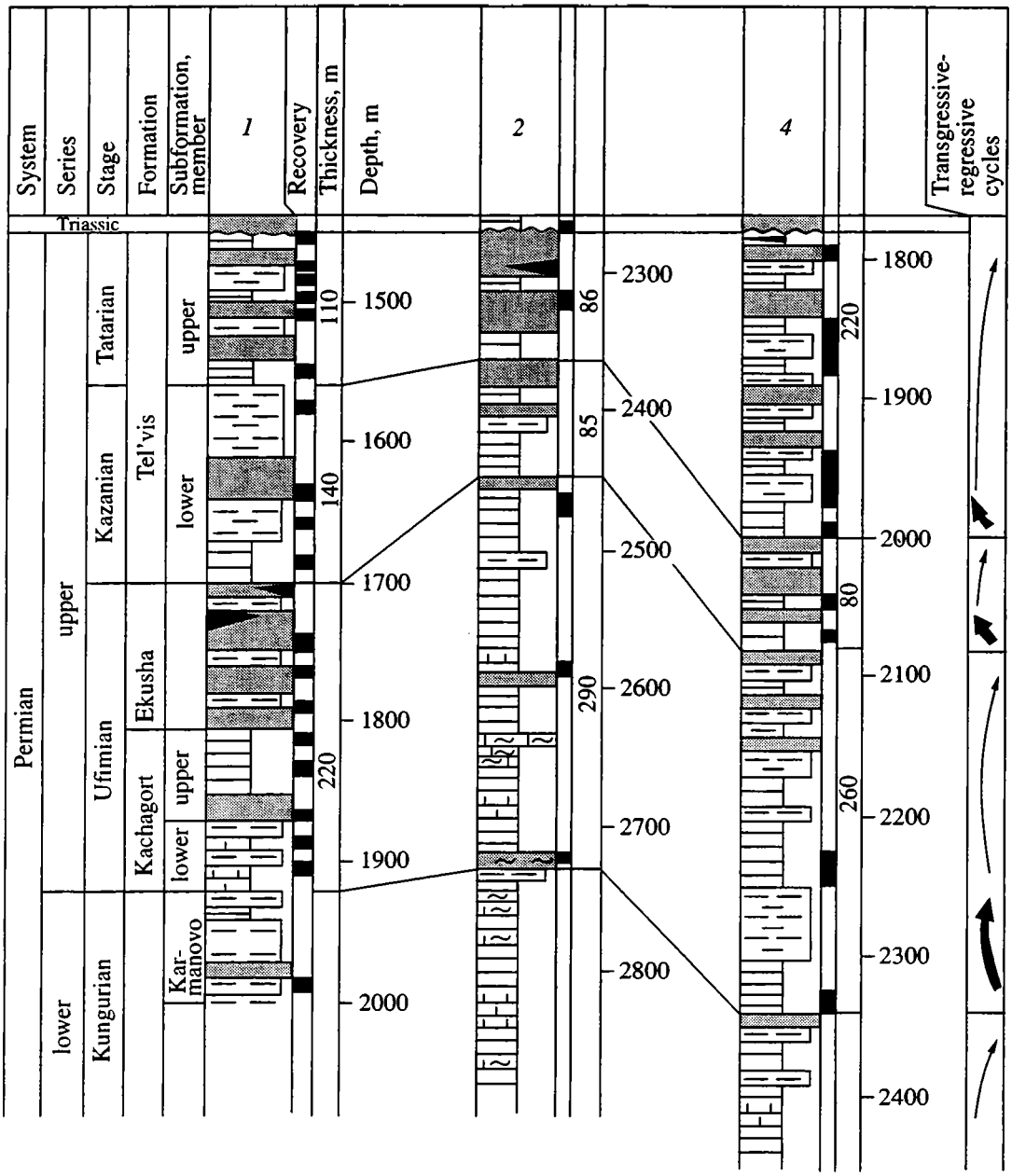

Fig. 3. Correlation of the Upper Permian sections penetrated by individual boreholes in Kolguev Island.
1 – Borehole 140 (after Entsova *et al.*, 1981), *2* – Borehole NW-202, *4* – Borehole PO-4.
For legend, see Figs. 1 and 2.

with sedimentary breccia interbeds. The deposits show both small- and large-scale cross-bedding, erosional surfaces at the smaller cycle boundaries, and traces of mud-flow processes. The sandstones are polymictic or, less commonly, quartzose in composition. Microquarzites and metamorphic micaceous schists dominate in the rock clasts; clasts of acid volcanic rocks occur as well. Interbeds of light quarzite-like sandstone appear in the upper part of the sequence. The age of this portion of the complex is based on a correlation with the Nibel' Formation in the Timan-Pechora region, which has a similar lithologic composition (Belyakova, 1988). The thickness is 880 m. The total thickness of the Ordovician sedimentary complex is more than 1500 m.

2. The Lower Devonian complex, which was penetrated by a single borehole (PO-46) in the Peschanoe Ozero field, is represented by a carbonate-terrigenous rock member overlying the green- and red-colored claystones of Middle (?) Ordovician age. The basal layer of the sequence consists of yellow-pink quartzose

sandstone with pebbles of quartz and quarzitelike sandstone. The thickness is 3 m. The middle part is composed of dolomites with dolomitized organogenic limestones. All of the rock varieties show an admixture of terrigenous quartz and anhydrite inclusions. Judging from the drilling sludge, the upper part of the section is made up of sandstone, siltstone, shale, and claystone. There are grey, green, and red rock varieties. The age of this complex is determined by the presence of conodonts *Pelekysgnathus* cf. *serratus* Ientzchs and *Pedavis* sp. in the limestone of the middle portion of the section (Borehole PO-46, depths 3736 and 3739 m), which are typical of the Lokhovian and Pragian stages. The thickness is 260 m.

3. The Upper Devonian complex is spread throughout the island. Usually, it transgressively overlies Lower-Middle Ordovician and, less commonly, Lower Devonian strata. The complex incorporates deposits of the Frasnian and Famennian stages, which can be subdivided into substages. A hiatus is recorded at the lower-upper Frasnian boundary. The upper part of the complex is eroded in most of island territory. The sequence is composed of terrigenous and, to a lesser extent, carbonate rocks; the latter are mainly present in the Famennian.

The lower part of the complex consists of coarse terrigenous rocks. In the northwestern, northern, and eastern areas of the island, this part of the section includes a thick lava flow of amygdaloidal basalt and tuff interbeds. In the northwest, the thickness of the flow is 75 m, whereas in the north and east, it decreases to 30 m. In the southern sequence basalts are absent. Beds of poorly sorted pinkish- and greenish-grey quartzose and polymictic sandstones with dispersed pebbles occur in the basal part of the section. The thickness of the beds ranges from 6 m to 15 m. A thick (more than 600 m) member of lilac-pink and cherry-brown sandstones, gravelstones, and conglomerates with individual boulders that are more than 10 cm in size is located only in the northern region of the island. Basic volcanic rocks, mica schists, quartzites, quartz, and red Ordovician rocks dominate the pebble composition. Acid and intermediate volcanic rocks and granitoids occur as well. The early Frasnian conodonts *Spathognathodus* gradatus (Younquist) and *Hindeodella germana* Hels. were determined in the middle part of the member (Borehole WP-1, depth 4350 m). The layers above the basalt flow contain early Frasnian floral remains of *Gigophyton gilkinefi* Lecl., *Archaeopteris acuta* Tschirk., and *Svalbardia* sp. and fishes *Asterolepsis* sp. and *Haplacanthus* sp. Claystones have lenticular and wavy bedding and contain pyrite and sometimes phosphate nodules and ooids. A member of calcareous claystones, marl, and clayey algal limestone occurs in the middle part of the section. Acritarchs and lingulid shells are present throughout the section. The basal layers yield an Upper Cambrian-lower Tremadocian acritarch assemblage, including *Ladogella longinsula* Golub et Volk., *L. rotundiformis* Golub et Volk. *Stellechinatum incinatum* Downie, *Stelliferidium* aff. *cortinullum* Denuff., *Cymatiogalea cuvilliera* Denuff., *Polygonum sexradiatum* (Tim.) Volk., *Acanthodiacrodium angustum* D.C., *A. polymorphum* Tim., *A. invictum* Rasul., *A. spinutisum* Tim., and others (Borehole NW-202, depths 4418 m and 4492 m).

The middle portion of the sequence shows the presence of the typical Tremadocian acritarch assemblage, which consists of *Saharidia fragilis* Downie, *Acanthodiacrodium comptulus* Rasul., *A.* aff. *ovatum* Tim., *Vulcanisphaera britanica* Rasul., *V. africana* Rasul., *Buedingsphaeridium tremadocum* Rasul., *Microhystridium shinetonensis* Downie, and others (Borehole NW-202, depth 4275 m; Borehole BG-1, depths 4150 m and 4082 m).

The late Tremadocian acritarch species *Pirea* sp. appears in the upper part of the section (Borehole NW-202, depth 4225 m). Lingulid shells of the species *Lingulinella* aff. *davisi* (MacCoy) occur throughout the sequence (Borehole NW-202, depth interval 4492 - 4225 m; Borehole BG-1, depth interval 4154 - 4082 m). This species is typical of the Upper Cambrian, but here it is recorded together with Tremadocian acritarchs. The thickness of the Tremadocian deposits exceeds 400 m.

The Arenigian part of the complex also represents a second- order cycle similar in structure to the Tremadocian one. The difference between them is that the former shows a predominance of red- and green-colored rocks over the grey-colored varieties, and that its siltstone and sandstone interbeds are greater in number and thickness. In the lower part of the cycle, Arenigian acritarchs *Baltisphaeridium hirsutoides* Eis., *B. brevispinosum* Eis., *B. multispinosum* Eis. (Borehole BG-1, depth 2802 m). In the northwest, this part of the complex sequence terminates with a red kaolinite-ferriferous (hematite-limonite) weathering crust that is overlain by an erosional contact with the upper Frasnian deposits. The thickness of the lower Frasnian portion of the complex varies from 10 m to 70 m (volcanic rocks excluded) in most of the distribution area; it exceeds 600 m only in the northern section.

The upper Frasnian portion of the complex is mainly composed of grey claystone, quartzose and polymictic siltstones, and sandstone with rare interbeds of algal limestone. Glauconitic varieties as well as rocks containing phosphatized and limonitized fossil remains occur. Amongst the latter, undeterminable brachiopods and conchostracans are present. In boreholes PO-4 and PO-46, these deposits yield a late Frasnian palynocomplex, which is subdivided into the Voronezh and Evlanovo subcomplexes (Dibner *et al.*, 1988). The thickness of the upper Frasnian deposits varies from 180 m in the south, to 400 m in the northwest, and to 230 - 280 m in the east of the island.

The Famennian portion of the complex in the most complete sequences (Boreholes PO-3 and NW-202) is subdivided into two substages, each representing a trasgressive-regressive cycle. The lower parts of the cycles

are composed of marine limestones, while the upper ones mainly consist of lagoon-marine and continental terrigenous rocks. A ferriferous weathering crust occurs at the top of the sequence.

The lower Famennian limestones are mostly represented by algal and oncolitic varieties. They are dolomitized to a different extent, and contain glauconite, pyrite, anhydrite, and barite. Brachiopods, echinoderms, early Famennian foraminifers (*Bisphaera irregularis* Bir., *B. gigantea* Con et Lis., and *B. malevkensis* Bir.), early *Famennian charoid* algae [*Umbella globula* Reitl., *U. regularis* Skr., *Quasiumbella rotunda* Byk., *Q. nana* (Reitl.), and *Pseudoumbella dondassica* Ber.], and other fossils occur throughout the sequence (Borehole PO-4, depths 3260 m and 3255 m). Limestone interbeds with ooid biogenic "bobolites" (*Bobolites petchoricus* Rad.) are also present (Borehole PO-4, depth 3240 m).

The upper portion of the Famennian cycle consists of quartzose and, less commonly, feldspar-quartzose sandstones, as well as siltstones, claystones, limestones, and dolomites with interbeds of the limonite and leptochlorite varieties. Grey, brown, greenish, and red-brown colors are typical of these deposits. Brachiopods, fish remains, and early Famennian miospores were reported from these strata (Dibner *et al.*, 1988).

Upper Famennian limestones predominating in the lower part of the substage section are represented by organogenic algal and briozoan-crinoid varieties, commonly with marl intercalations. Late Famennian foraminifers occur throughout the sequence, the lower part of which yields *Septaglomospiranella nana* Reitl., *S. parva* (Durk.), *S. romanica* Lip., *Septatournayella rauserae* Lip., *Septabrunsiina kingirica* (Reitl.), *Tournayella discoidea* Dain., and others (Borehole PO-4, depth 3132 m; Borehole PO-8, depth 3329 m). The foraminiferal assemblage from the top layers of the limestone is dominated by *Quasiendothyra communis* (Raus.), *Q.* ex gr. *konensis* Leba, and *Q. kobeitusana* (Raus.) (Borehole NW-202, depth 3700 m; Borehole WP-1, depth 3840 m). Marl and claystone interbeds at the latter levels also yield the upper Famennian miospore assemblage.

The upper layers of the upper Famennian Substage are composed of terrigenous rocks: dark-grey claystones and clays with interlayers of pink, greenish-grey, and white quartzose sandstones and siltstones. The top part of the sequence shows the presence of red claystones grading into red-brown and dark-blue-black hematitic rocks, which most likely represent a weathering crust. Grey clayey beds contain a miospore assemblage with *Retispora lepidophyta* (Kedo) (Borehole PO-8, depth 3211 m).

The thickness of the Famennian deposits of the Upper Devonian sedimentary complex varies from 50 m to 340 m. Its most complete sequences occur in the northwestern and northern areas as well as in Borehole PO-8 in the Peschanoe Ozero field. The total thickness of the Upper Devonian rocks ranges from 320 m in the south, to 560 m in the northwest, and to 620 m in the east of the island. Their maximum thickness (more than 1040 m) is recorded in the northern part of the island (Borehole WP-1).

4. The Lower Carboniferous sedimentary complex is subdivided into two subcomplexes: the Tournaisian and the upper Visean-Serpukhovian. The former occurs in the northern part of the island, whereas the latter is spread throughout the whole area.

The Tournaisian part of the complex begins with a carbonate-terrigenous variegated member (about 20 m thick) of lagoon-continental genesis, which yielded the early Tournesian miospores *Vallatisporites pusillites* (Kedo) Dolby et Neves and *Verrucosisporites gobbettii* Pleford (Borehole PO-8, depth 3204 m). This member rests with an erosional contact upon the ferriferous weathering crust. Higher Tournaisian layers are mainly composed of limestones and, less commonly, of dolomites with interbeds of quartzose sandstones, clayey siltstones, and claystones. Grey, dark-grey, and black colors are typical of these rocks; they are enriched in organic matter and contain pyrite. Late Tournaisian foraminifers *Chernyshinella glomiformis* and *Ch. (Rectochernyshinella) mirabilis* (Lip.) were determined in the limestone beds (Borehole WP-1, depth 3663 m). The thickness of the Tournaisian rocks is 120 m and 9 m in the northern section and Peschanoe Ozero field, respectively.

Upper Visean deposits are composed throughout the area of light organogenic foraminiferal-crinoid limestones and brownish-grey to brown dolomites with interbeds of white anhydrite and dark-grey claystone enriched in organic matter. There are beds within the limestone members that are completely composed of shells of the brachiopod species *Gigantoproductus* ex gr. *moderatus* Schw., *G. latiexpansus* Sar., and less common *Striatifera striata* Fisch. (Borehole NW-202, depth 3595 m). This facies is typical of the upper Visean in the Russian Platform. Based on foraminifers, the Aleksin, Mikhailov, and Venev horizons were established in the upper Visean sequence. The first one is characterized by *Pseudoendothyra struvei* Moeller, *P.* cf. *subpressa* Schlijk., *Archaeosiscus krestovnikovi* Raus., *A. karreri* Brody, *Endothyranopsis intermedia* Raus., and others (Borehole PO-3, depth interval 3507 - 3457 m; Borehole NW-202, depth 3595 m). The Mikhailov Horizon contains abundant *Globoendothyra*; in addition, the first occurrences of *Neoarchaediscus rugusus* Raus. and *Asteroarchaediscus* cf. *ovoides* Raus. are recorded here (Borehole PO-3, depth interval 3457-3452 m). In the top layers of the horizon, the assemblage composition becomes impoverished, which is related to dolomite predominance. For this reason, the foraminiferal assemblage in the Venev Horizon is dominated by various *Eostaffella* (*E. ikensis* Viss., *E. parasrtuvei* Raus., and others; Borehole PO-3, depth interval 3452 - 3448 m; Borehole NW-202, depth

3560 m). The thickness of the upper Visean rocks varies from 20 m in the south to 60 m in the east of the island.

The Serpukhovian sequence is lithologically dominated by white, bluish-grey, and pink anhydrite and dark-brown dolomite with celestine. There are also interlayers of carbonate breccia, bluish-green claystone, and pinkish-brown and bluish quartzose sandstone. Foraminifers occur in the upper part of the sequence, where limestone interbeds are present. Their most diverse assemblage, with *Omphalotis geniculata* Gan., *O. minima* Raus., *Eostaffelina delicata* Ros., *E. paraprotvae* Raus., *E. protvae* Raus., *Eostaffella parastruvei*, *E. proikensis* Raus., and others, was recorded in the eastern part of Kolguev Island (Borehole PO-4, depth 3005 m). The thickness of the Serpukhovian deposits varies from 30 m to 50 - 120 m in the south and east, respectively. In the southern and northwestern areas, drilling sludge yielded foraminifers *Millerella carbonica* Grozd. et Leb., *M. paracarbonica* Man., *Pseudostaffella antiqua* (Dutk.), and *Ozawainella* aff. *umbonata* Brazhn. et Pot. (Borehole BG-1, depth 2526 m; Borehole NW-202, depth 3474 m), which are characteristic of the Bashkirian Stage. The complex probably also included a part of the Bashkirian Stage, which was composed of organogenic limestones and later destroyed by pre-Moscovian erosion in most of the area.

The thickness of the complex ranges from 60 m in the south, to 100 - 180 m in the east, and 340 m in the north of the island.

5. The Middle Carboniferous-Sakmarian complex usually begins with the Podol'sk Horizon of the upper Moscovian Substage; only in the northwestern area are its basal beds dated by the Kashirian Stage. Almost the entire Bashkirian Stage and the lower Moscovian Substage are missing in the section throughout most of the area. A contact between the Serpukhovian and upper Moscovian strata was not observed; however, in most boreholes, finds of fusulinids with the respective ages are spaced so closely that there is no room for any basal terrigenous member. None of its traces were recorded by logging either, according to which the Moscovian-Serpukhovian boundary corresponds to the appearance of dense organogenic limestones.

The lower and upper parts of the complex are essentially different. The lower part, comprising the upper Moscovian Substage and the Upper Carboniferous, is, in general, laterally stable in thickness. The upper, Asselian-Sakmarian part occurs as a chain of reefs with essential facies diversity in different areas. Therefore, these deposits will be considered separately.

The Middle-Upper Carboniferous deposits are dominated by grey, sometimes almost white organogenic limestones composed of brachiopod, briozoan, and crinoid remains. In the northwestern and northeastern sections, the Upper Carboniferous is marked by the appearance of greenish-grey and dark-grey marls and by a change in the limestone color, which becomes darker. Traces of sediment slumping and inclusions of white organogenic limestone in marls and black limestones are typical features of these deposits.

The Middle-Upper Carboniferous rocks are highly silicified. The siliceous material is crystallized to various extents and occurs as both an unevenly dispersed admixture in limestones and lenses, or intercalations with a thickness range of 10 - 15 cm. The Middle Carboniferous limestones show the presence of anhydrite inclusions, whereas the Upper Carboniferous rocks contain glauconite grains. Paleontologically, the Middle and Upper Carboniferous deposits essentially differ from each other.

The Middle Carboniferous portion of the section contains diverse fusulinid assemblages typical of the following units: the Kashira Horizon with *Neostaffella subquadrata* (Grozd. et Leb.), *N. vozhgalica* (Saf.), *N. kremsi* (Raus.), *Profusulinella meridiana* Bensh., *Ozawainella mosquensis* Raus., *O. kumpani* Sosn., and *O. paratingi* Man (Borehole NW-202, depth 3405 m); the Podol'sk Horizon with *Parawedekindellina subovata* Saf., *P. kamensis* Saf., *P. curta* Grozd., *Pseudostaffella* cf. *ivanovi* Raus. and others (Borehole BG-1, depth 2500 m); and the Myachkovo Horizon with *Fusulinella laxa* Sheng, *Wedekindellina excentrica* Roth. et Sc., *Schubertella acuta* Raus., and others (Borehole BG-1, depth 2422 m). At the Middle-Upper Carboniferous boundary, the last occurrences of *Wedekindellina* forms are recorded together with the presence of *Protriticites* species (*P. pseudomontiparus* Put. and *P. ovatum* Put.) (Borehole BG-1, depth 2420 m). Unlike the Middle Carboniferous, the basal Upper Carboniferous beds are almost barren of fusulinids, although their presence in a minimal volume (about 20 m) is substantiated by the find of Kasimovian conodonts *Streptognatodus excelsus* St. et Pl., *S. cancellosum* (Gummel.), and *S.* cf. *gracilis* St. et Pl. (Borehole BG-1, depth 2398 m). The Gzhelian strata show the presence of fusulinids and smaller foraminifers throughout the sequence. Its lower portion is characterized by *Pseudofusulinella prima* Skinner et Wild., *P. annae* (Grozd.), *P. pulchra* Raus. et Bel., *P.* aff. *usvae* Dutk., *P. bella* (Grozd. et leb.), and *Fusiella* ex gr. *schubertellinoides* Sul. (Borehole WP-1, depth 3182 m), whereas the upper one contians *Pseudofusulinella antiqua* Skin. et Wild., *P. delicata* Skin. et Wild., *Schwellwienia modesta* (Scherb.), *Sch. cognata* (Konov.), *Sch. signata* (Kon.), and *Schwagerina lilia* (Kon.) (Borehole NW-202, depth 3225 m).

The Upper Moscovian and Upper Carboniferous brachiopods compose a mutual assemblage, which lacks choristites typical of the Russian platform and the Timan-Pechora province. The assemblage is similar to that from the Zilim Beds in the Urals region. Occasional finds of well-preserved long spines of brachiopod shells suggests that their habitat was located at considerable depths, where wave influence was insignifi-

cant. The most common forms are *Plicatifera pseudoplicatilis* Step., *Jakovlevia pseudoartiensis* Stuck., *Achunoproductus achunowensis* Step., *Eomarginifera timanica* Tschern., and *Brachythyrina stranwaysi* Vern. (Borehole BG-1, depths 2500, 2419, and 2398 m). In the northeastern area, the brachiopod assemblage becomes more uniform, and the shells decrease in size, indicating basin deepening.

The thickness of the upper Moscovian deposits varies from 80 to 140 m. It has minimal values in the south, near the Bugrino Settlement, and gradually increases northward and northeastward. Variations in the thickness of the Upper Carboniferous sections are less significant, from 60 to 80 m; this value decreases northeastward.

Rocks corresponding to three facies can be distinguished in the upper part of the complex.

In most boreholes drilled in the Peschanoe Ozero field in the eastern part of the island, Asselian-Sakmarian deposits are represented by grey and greenish-grey clayey limestones and marls showing sediment slumping structures and, sometimes, the presence of pyrite. Lithologically, this section resembles that of the Sezym Formation in the Pechora basin. Fusulinids are relatively rare, and they mainly occur in the Asselian basal layers (*Pseudofusulinella nana* Semina, *P.* cf. *blochini* Korzh., *Daixina* cf. *tschernovi* Mukh., and *Triticites arcticus* Schellw.) (Borehole PO-4, depth interval 2788 - 2816 m). In addition to these species, an assemblage of small nodosariids (*Protonodosaria praecursor* Raus., *Nodosaria clavatoides* Raus., *N. bella* Lip., *N. parva* Lip., and others), which are typical of the Sezym Formation, appears here. Brachiopods show a supressed aspect, similar to their species from the Upper Carboniferous strata, which hinders an identification of the Carboniferous-Permian boundary.

A 6-km-wide band of reefs and their clastic material occupy the western Pechanoe Ozero field and then turn to the northwest (Preobrazhenskaya *et al.*, 1993). The rocks are represented by light massive limestones composed of various algae (Charophyta, Rhodophyta, and Cyanophyta), sometimes associated with bryozoans. In addition to the reef rocks themselves, the cores display their destruction products, such as breccia composed of small reefal clasts, usually strongly porous and weakly cemented. The reefal rock complex comprises both the Asselian and the Sakmarian stages (at least the Tastub Horizon), as is implied by the fusulinid assemblage of *Schubertella transitoria* Staff., *Sch.* ex gr. *paramelonica* Sul., *Sch. sphaerica* Raus., *Triticites* ex gr. *shikhanensis* Kosov., and *T.* ex gr. *schwageriniformis* Raus. (Borehole PO-38, depth 2906 m).

The third type in the section was drilled in the Tarka field in the center of the island. Its main feature is the presence of light-grey, sometimes almost white anhydrite, which is absent in other sections. The dominant rocks are dark-grey laminated clayey limestones that are enriched in bryozoan remains, commonly dolomitized, and show intercalations of light-brown dolomites. Well-preserved sponges and plant debris occur in these rocks. The synchronization of these rocks with the reefal deposits proper is substantiated by rare finds of the fusulinid species *Schubertella sphaerica* and *Sch. compacta* Sul., although the foraminiferal assemblage is dominated by nodosariids *Protonodosaria subcuboides* (Raus.), *P. subtensis* (Grozd. et Leb.), and others (Borehole T-1, depth 2998 m; Borehole T-7, depth 3080 m).

The gypsum-bearing deposits of the considered area evidently represent a back-reef lagoon complex. If this is true, then the sediment of the first zone should be considered as having been deposited in the reef-slope environments of the transitional zone between the Novaya Zemlya and southern Barents deep basins.

The thickness of the Asselian-Sakmarian sequence ranges from 50 m to 250 m. The maximum values are recorded in the central part of the island, i.e., in the lagoonal zone, while in the reefal area, the thickness decreases to 160 - 180 m. Eastward, it continues to decrease to common values of 110 m and, sometimes, 50 m. This most likely is caused by the partial eastward and northeastward transportation of the material by slumps into a transitional zone of uncompensated sedimentation beyond the limits of the island.

The total thickness of the complex somewhat increases in the southeastern direction (from 280 to 320 m). In the extreme eastern sequence, it again decreases to 200 m. The maximum thickness (480 m) is recorded in the northwestern area of the island.

6. The Artinskian-Tatarian sedimentary complex is mainly composed of terrigenous rocks. Thick limestone members are observed only in the Artinskian deposits. The transition from the carbonate deposits of the previous complex to the terrigenous rocks is gradual; therefore, the boundary between these two complexes is unclear. It is tentatively drawn at the base of the first thick terrigenous member.

In general, these rocks are similar to coeval deposits in the Timan-Pechora region, though they differ from them in their predominance of marine rock varieties and smaller amount of sandstones. For instance, the thickest sandstone member, which was previously defined in the Timan-Pechora region and the southern part of Kolguev Island as the Ekusha Formation (Entsova *et al.*, 1981), sharply decreases in thickness in other parts of the island, until it is completely wedged out in the northwest. This hinders the usage of formational subdivision based on the Bugrino section for the entire territory of the island (Fig. 3).

The thick terrigenous member has a cyclic structure. The lower transgressive part of the cycles is mainly composed of claystones bearing marine faunal fossils and associated with marls (in the lower portion of the complex). Upsequence, the amount and dimensions of the terrigenous material increases, and lagoon and continental facies appear. The cycles terminate with essen-

tially sandy members with plant remains, coaly intercalations, and sometimes coal-bearing beds. The distinguished major cycles served as a basis for subdividing and correlating the sequence. Since the direct correlation of these units with the stratotypes of the Upper Permian stages is impossible because of essential differences in their paleontologic characteristics, the cycle boundaries were tentatively considered to represent the stage boundaries (Fig. 3). This tolerably corresponds to the palynological data, which allows the sequence to be correlated with those from the well-studied Pechora basin.

The changes in the estimates of the deposit ages, as compared with the previous determination (Entsova *et al.*, 1981), are the following: (1) the entire Karmanovo Member of the Bugrino section, the lower part of which was considered to be still of Artinskian age, is referred to the Kungurian Stage; (2) the Kachgort and Ekusha formations are referred to the Ufimian Stage, although the latter was previously considered to be a part of the Kazanian Stage; and (3) we refer the upper Tel'vis Formation to the Tatarian Stage, although its main part was previously thought to be of the Kazanian age.

These new datings, which resulted from the paleontological analysis, correspond well to the latest data from other regions. For instance, a significant part of the deposits, which for a long time was correlated in the Arctic region with the Kazanian, is now referred to either the Ufimian or Tatarian stages (*Osnovnye...*, 1984).

The Artinskian sequence consists of limestones and terrigenous rocks, sometimes with a considerable silica admixture. The rock proportions vary throughout the area. The limestones are grey, sometimes pinkish, organogenic, and commonly contain chert inclusions and abundant briozoan and crinoid fragments. The claystones are dark-grey and, less commonly, red-brown in color. The siltstones and sandstones are light-grey, quartzose in composition, and show spongolite inclusions. The rocks contain the typical Artinskian brachiopod assemblage, including *Anidanthus ufensis* Stuck., *Uraloproductus stuckenbergianus* Krot., *Jakovlevia mammatiformis* Pred., and *Spiriferella saranae* Vern. (Borehole T-1, depth interval 2795-2799 m). The basal layers of the sequence in the northwestern area yielded the early Artinskian (Burtsevka Horizon) fusulinid assemblage of *Waeringella schoriensis* Kon., *Pseudofusulina trivialis* Shir., *P. arlarovi* Raus., *P. pedisequa* Viss., *P. concessa* Viss., *P. adjuncta* Shir., and others (Borehole NW-202, depth 3094 m). In addition to the typical Sakmarian protonodosariids, the smaller foraminifers include species *Nodosaria cassiaformis* Igon., *N. bogatirevi* Igon., *N. incelebrata* Gerke, *Frondicularia prima* Gerke, *F. tigianica* Gerke, and *F. mica* Gerke. These species are usually considered to be characteristic Kungurian forms, but they also appear at the base of the Artinskian Stage (Borehole T-1, depth interval 2795 - 2799 m).

The thickness ranges from 85 to 180 m. The smallest thickness was recorded in the center, whereas the maximum values occur in the south and northwest of the island.

The Kungurian sequence is dominated by dark-grey and grey claystones and clayey siltstones, usually silicified and enclosing marl and, less commonly, limestone interbeds. Upsection, quartzose siltstones and sandstones dominate. Macrofossils are rare and represented by bivalves and gastropods, while the rare carbonate interbeds are enriched in foraminifers. The latter are mostly agglutinated forms of the *Psammosphaera*, *Ammodiscus*, *Hyperammina*, and *Hyperamminoides* genera (Borehole PO-3, depth 2940 m), which are typical of the horizon with agglutinated foraminifers defined by Gerke in the Nordwick area. Higher, calcareous forms dominate. In addition to protonodosariids, typical Kungurian *Nodosaria cassiaformis* Igon., *Frondicularia prima* Gerke, *F. amigdaleformis* Gerke, and other species are present, associated with *Nodosaria hexagona* Tscherd., *N. pseudoincelebrata* Sossip., and other forms, which are usually considered to be younger in age (Borehole PO-4, depth interval 2477 - 2513 m; Borehole T-1, depth 2700 m).

The thickness of the Kungurian sequences varies from 110 to 200 m. Minimal thicknesses are characteristic of the southern and central parts of the island, whereas the thickest deposits were observed in its northwestern and eastern areas.

The Kachgort and Ekusha formations of the Bugrino section and their analogues in other areas, where the thickness of the latter formation sharply decreases, are referred to the Ufimian Stage. The deposits exemplify a distinct sedimentary cycle, which begins with a member of claystone, marl, and limestone or glauconite polymictic sandstone. Upsection, these rocks are replaced by claystone and a clayey siltstone member. The cycle terminates with a paralic member of sandstones with coal (in the south) and coastal-marine or deltaic sandstone beds (in the east). The sandstones are greenish-grey, fine-grained, and sometimes weakly lithified. The clasts are mostly composed of acid and basic volcanic rocks; clasts of metamorphic schist occur as well. In the northwest, the Ufimian deposits show such features as a sharp decrease in the number of sandstone interbeds, considerable silica contents in clayey or silty rock varieties, the presence of spongolite intercalations and inclusions, and a high content of glauconite and sometimes tuffogenic material.

The diverse marine fauna includes forms characteristic of both the Kungurian and Kazanian stages. Amongst the smaller foraminifers, there are abundant typical Upper Permian *Pseudonodosaria lata* A.M.-Maclay, *Rectoglandulina pigmaeformis* A.M.-Maclay, *Frondicularia dilemma* Gerke, *F. heminflata* Gerke, *F. reliqua* Gerke, and others, which cooccur with protonodosariids characteristic of the underlying strata (Boreholes T-1 and T-7, depth interval

2560 - 2588 m; Borehole PO-3, depths 2652 m and 2540 m).

Brachiopods are represented by the Svalbard assemblage, which includes *Megousia kulikii* Fred., *Anidanthus aagardi* Toula, *Jakovlevia mammatiformis* Fred., *Anemonaria variabilis* Ustr., *Svalbardoproductus* sp., *Heteralosia* ex gr. *excavata* Gein., *Spiriferella keilhavi* Ruch., and others (Boreholes T-1, T-7, and PO-3, depth interval 2560-2588 m and 2540 m). Although rare Artinskian forms are present in the briozoan assemblage, it is dominated by species and even genera (*Parafenestralia* and *Primorella*; Borehole T-1, depth 2560 m) that, so far, are known only from the Upper Permian. In addition, common fragments of inoceramlike bivalve shells, which are widespread throughout the Upper Permian, appear there. These facts allow the Late Permian age to be undoubtedly assigned to the considered part of the section, and they make the Lower-Upper Remain boundary less ambiguous.

The thickness of the Ufimian section shows less variations than that of the Lower Permian strata. The minimal thickness of the Ufimian deposits (220 m) occurs in the south, whereas northward and eastward, it increases to 280 - 290 m.

We refer the Tel'vis Formation of the Bugrino area and its analogues to the Kazanian and Tatarian stages. The two cycles distinguishable in this portion of the section correspond, judging from palynospectra, to the Kazanian and Tatarian stages, respectively. The number and thicknesses of the sandstone beds noticeably increase in the top unit of the cycle. The sorting of material in this part of the section is considerably poorer than in the underlying strata, which is related to the wider occurrence of lagoon facies.

The macrofossil abundance sharply decreases here. Amongst brachiopods, *Megousia weyprechti* Toula, *Spitzbergenia alferovi* Milor (Borehole PO-22, depth 2260.5 m), and large *Lingula freboldi* Gobbett. (Borehole PO-1, depth interval 1942 - 1944 m) are the most common. They form thin coquina interbeds in the bar facies. The assemblage lacks the forms typical of the Kazanian Stage in the Russian platform. Bivalves are also common, in particular, inoceramlike *Atomodesma* sp. (Borehole NW-202, depth 2404 m).

Smaller foraminifers, whose assemblage (more than 30 species) is believed to be characteristic of the Kazanian Stage in the Arctic region, are abundant and diverse. Most common among them are *Nodosaria solidissima* Gerke, *N. gigantea* Sossip., *Lingulinella arctica* Gerke, *Rectoglandulina borealis* Gerke, *R. pigmaeformis* A.M.-Maclay, *Frondicularia dilemma* Gerke, *F. reliqua* Gerke, *F. longissima* A.M.-Maclay, *Tristix permiana* Gerke, and others (Borehole PO-3, depth 2498 m; Borehole T-1, depth 2243 m). In the upper portion of the sequence, almost all genera include new species. Further study of these species will probably allow the characteristic Tatarian assemblage to be defined, although the uppermost 30 - 40 m of the section are barren of foraminifers. These strata contain only floral remains of *Cordaites*(?) cf. *iljinskiensis* (Radcz.) S. Meyen, *C. concinnus* (Radcz.) *S. Meyen*, and *Rufloria*(?) cf. *minuta* (Radcz.) (Borehole T-4, depth 2035 m) and freshwater bivalves *Soanella*(?) *linguliformis* (Lutk.) and *Adzvaella tatarica* (Kanev.) (Borehole T-4, depth interval 2028 - 2032 m). Based on the palynospectra, the considered part of the section can be subdivided into the Kazanian Stage and the lower and upper Tatarian substages. The continental deposits of the upper Tatarian Substage demonstrate a transition from the small-scale "sandstone-claystone" cycles characteristic of the coal-bearing continental beds to the "claystone-sandstone" cycles widespread in the underlying, predominantly marine strata.

The thickness of the Kazanian section commonly varies from 80 m to 140 m, and it increases to 216 m only in the extreme eastern area (Borehole PO-3). The thickness of the Tatarian deposits displays more significant variations. Its values range from 110 m in the south and northwest of Kolguev Island, to 180 m in its central part, and then to 240 m in the Peschanoe Ozero field.

The total thickness of this complex regularly increases from 760 m in the south, to 790 m in the Tarka area in the central part of the island, and to 1080 m in the Peschanoe Ozero field.

Despite traces of local erosion of the upper Tatarian strata, a considerable hiatus seems to be lacking at the Permian-Triassic boundary.

CONCLUSION

(1) As a result of lithologic-stratigraphic studies by a team of scientists, Kolguev Island has become one of the best studied areas in the Timan-Pechora province. The defined lithologic complexes, their datings, and the estimated hiatus spans between these complexes serve as a reliable basis for interpreting the seismic data in adjacent areas.

(2) The Lower-Middle Paleozoic sequence of Kolguev Island is typical of young platforms. Distinct facial and thickness variations along with the presence of Ordovician and Devonian conglomerates and basalts testify to the occurrence of intense block movements during those times. Typical platform facies common in the Russian platform, such as organogenic limestones consisting of *Gigantoproductus* and *Striatifera* spread throughout the island during the Lower Carboniferous. The thickness of the Devonian-Lower Carboniferous deposits (data on the Ordovician are insufficient) is minimal (380 m) in the south of the island. It increases to 660 m northwestward and to 800 - 1000 m eastward, toward the Pai Khoi-Novaya Zemlya basin. Their thickness increases even more sharply (to 1380 m) northward, toward the southern Barents basin, which may indicate a much older age for this structure, despite the widely shared opinion about its Late Paleozoic origin (*Barentsevskaya...*, 1988).

(3) In contrast to the relative uniformity of the Lower-Middle Carboniferous sequences throughout the area, a new type of facies zonality appears in the Late Carboniferous, and it becomes most pronounced in the Asselian-Sakmarian. A chain of reefs, which stretched in the form of a northward-curving arc almost throughout the island, appeared at that time. Inside the arc, lagoonal anhydrite-bearing facies were abundant. To the north and northeast of it, the slope facies of sediments with slumping structures, similar to those in the Sezym Formation of the Pechora basin, were deposited. In the Pai Khoi and Novaya Zemlya region, the Sezym Formation is replaced by abyssal condensed sequences of Mn-rich siliceous-clayey shales (Ustritskii, 1985). It is hardly improbable that a similar replacement also occurs to the north of Kolguev Island, in the southern Barents basin, where the Upper Paleozoic sequence is believed, by analogy with the Timan-Pechora region, to be composed of carbonate rocks (*Barentsevskaya...*, 1988).

(4) During the Artinskian, carbonate sedimentation gave way to deposition of terrigenous sedimentation, owing to the initiation of an influx of clastic material from the Urals. The sequences again become rather uniform, although they differ from those in most of the Timan-Pechora region in the decreased number and thickness of their sandstone members and their increased share of marine deposits. Therefore, formations that are traceable throughout most of the Timan-Pechora region are recognizable only in the Bugrino sequence, in the south of the island. Northeastward, the thickness of the Artinskian-Upper Permian deposits also increases again by one-half (from 760 m in the south to 1080 m in the Peschanoe Ozero field), which agrees well with the existence of the Early Permian deeper marine basin in the north and east of the island.

(5) The modern position of Kolguev Island at the Pechora Sea shelf is closely related to the Late Paleozoic structure of the region. In the Early to Middle Carboniferous, a uniform shallow marine basin existed in the entire Timan-Pechora region, comprising its modern sea continuation. Later, at the Middle-Late Carboniferous boundary, the northeastern and northwestern areas of the region experienced subsidence, and, the shelf-break zone of the Pai Khoi-Novaaya Zemlya basin (ocean?) moved away from the southwestern Pai Khoi and South Island of the Novaya Zemlya Archipelago toward Kolguev Island for a distance of 250 km. The part of the shelf that was located between Kolguev Island and the southern Barents deep basin also experienced a similar subsidence. Thus, Kolguev Island appears to be situated in the northeastern corner of the Late Paleozoic platform of the Timan-Pechora region. During the Late Paleozoic, the areas to the northeast and northwest of the island represented a transitional zone to the southern Barents and Pai Khoi-Novaya Zemlya deep basins.

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Late Jurassic Benthic Assemblages of Siberia

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Abstract – Four shallow-water benthic assemblages are described. They populated soft ground in the marine gulf in the northern Taimyr Peninsula during the Late Jurassic (middle Volgian, the *Maximus* Zone). The successive chronological changes in these assemblages are considered to be a primary ecological succession related to settling in newly formed biotopes. A comparative study of the evolving assemblages shows that both the quantitative and qualitative changes in them fit those of the modern ecological successions. Thus, the paleosuccessions may serve as models for a historical forecast of recent ecological successions caused by both natural catastrophes and human activity.

Key words: **Benthic assemblages, Late Jurassic, ecological successions, Taimyr Peninsula.**

Ecological successions are one of the basic, natural phenomena. They are well-studied for land conditions, for both water deficiency (xeric successions) and redundancy (hydric successions) (Ricklefs, 1976). However, successions in marine environments have been less studied. In paleontology, this phenomenon does not receive proper attention. International projects (IGCP projects 216 and 293) that focus on biosphere evolution have mainly been aimed at studying crisis events in biota development (*Global Bio-Events...*, 1986; *Extinction Events...*, 1990). Only the current project "Biotic Recoveries from Mass Extinctions" (IGCP projects 335) places emphasis on analyzing communities and biotas after a crisis. These studies are not only oriented toward communities and ecosystems during a crisis event, but they also focus on assemblages that survive. This includes the documentation of taxonomic and morphological diversity, the pattern and rate of adaptation to the environment, and the dynamics of the main features of communities and ecosystems. During the decade-long activity of the previous project 216 "Global Biological Events in Earth History," headed by O. Walliser, the above-mentioned problem was also addressed. However, in the current project, it is suggested that attention be shifted from the question "What has happened?" to the question "What are the consequences?". The purpose is to better understand the ways that ecosystems are restored after their destruction in the geological past in order to take effective measures for the accelerated restoration of damaged ecosystems during historical periods (D. Erwin, E. Kauffmann; First Circular, August, 1992).

In my opinion, the outlined idea justifies one of the directions in studying the restoration of communities and ecosystems. The point in question is a series of shallow marine benthic communities that occupied the marine basin formed in the northern Taimyr Peninsula at the end of the Jurassic (Fig. 1).

Material. A small bay in the northern Taimyr Peninsula was formed as a result of the major boreal transgression in northern Eurasia during the middle Volgian (Fig. 1). Here, in the Leningradskaya River basin, a terrigenous sequence was deposited over a short period (a part of the *Dorsoplanites maximus* Zone). The deposits are mainly represented by sandy sediments with an observed thickness of up to 90 m (Fig. 2). Rocks of the *Maximus* zone lie directly upon Silurian shales and are covered by Quaternary deposits (Basov *et al.*, 1965). The sediment composition and section structure suggest the following scenario of basin evolution.

During the initial stage, the basin was represented by a shallow lagoon, which was almost isolated from the sea and, apparently, freshened (deposition time of Beds 1 - 2). The extreme diversity and instability of the substrate, caused by the active transportation of unsorted fragments of black shales from the surrounding land, hampered the formation of stable populations of invertebrates.

During the second stage (formation time of Beds 3 - 7), the lagoon became more closely connected with the sea. This resulted in the origination of a shallow and, apparently, still freshened bay. Coarse clastic substrates were predominant, but the facies were more stable and uniform. This is indicated by both allochthonous and autochthonous fossils: bivalves, brachiopods, foraminifers, and ichnofossils of invertebrates.

During the third stage (deposition time of Beds 8 - 19), the area became a marginal shallow marine zone, more distant from the coast. Fine-grained and, sometimes, muddy substrates became predominant. They were inhabited by numerous benthic communities, which were alternatively dominated by oysters, arctotises, or large pectinids.

During the last stage (Beds 16 - 20), diversity reached a maximum. The assemblage included dozens

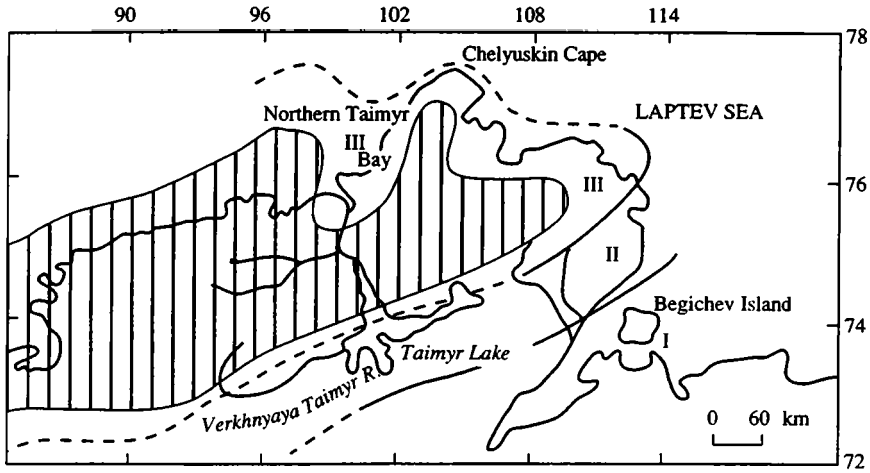

Fig. 1. Paleogeography of the North in the Late Jurassic (middle Volgian time), the *Dorsoplanites maximus* Zone. Marine conditions: I – deep, II – moderately deep, III – shallow. Land areas are hatched.

of species of bivalves, brachiopods, gastropods, and worms. The microbenthos was represented by numerous foraminifers and ostracodes. During the last stage of the bay formation, semipelagic forms were rare (there are finds of ammonite shells and rostra of belemnites).

Research methods. The term "succession" is used in marine hydrobiology to describe the sequential settling of hydrobionts in newly formed biotopes, for example, the surface of submarine lava flows (Oshurkov and Ivanyushina, 1991). In the paleoecology of land plants, it has been suggested that a change in communities over time be described as a paleosuccession (Krassilov, 1972). As for terminology, the author shares the opinion of those paleontologists who use neontological concepts and terms without major reservations, being aware of an evident difference in the content and quality of information concerning the recent and fossil taxa and taxonomic groups.

The present article is devoted to the analysis of temporal changes in the marine benthic communities. These communities were reconstructed on the basis of an established sequence of autochthonous elements of thanatocoenoses in the *Maximus* Zone of the Volgian Stage (Upper Jurassic) (Zakharov, 1981, pp. 173 - 177). All available lithologic and taphonomic observations were used in the course of the reconstruction of communities (Fagerstrom, 1964; Zakharov, 1975, 1988).

The functional unit of a community is represented by the adaptive form (*forma accomodata*). It comprises taxa with a similar mode of life. The adaptive form of the Mesozoic bivalves corresponds to a taxon of the genus rank.

The quantitative estimation of various genera in oryctocoenoses was carried out using the method that was previously elaborated by the author (*Opornyi Razrez...*, 1969; Zakharov, 1981). It allows for a rough estimation of taxa occurrence on a seven-point scale: 100 (abundant) – finds of hundreds of specimens, 30 (very many) – several dozens of specimens, 9 (many) – a few dozens, 5 (very frequent) – finds of 11 - 15 specimens, 3 (frequent) – finds of 6 - 10 specimens, 2 (rare) – from 3 to 5 specimens, and 1 (very rare) – occurrence of 1 to 2 specimens. Points 100, 30, 9, 5, 3, 2, and 1 show the conditional ratios of a group in the oryctocoenosis. Because of this, they are referred to as abundance indices. Points were used for the construction of histograms and for percentage calculations in circular diagrams.

RESULTS

1. Pioneer community. Stage of first settlers. The benthic community most typical for the first stage is reconstructed for the oryctocoenosis found in Bed 4 (Fig. 3a). It includes three sessile suspension feeders of the upper level (oyster, buchia, and terebratulid) and one genus of vagile suspension feeders of the lower level (*Astarte*). The pioneer community is distinct in the following: (1) low taxonomic diversity, (2) well-levelled ratio of specimens, (3) absence of a marked predominance in combination with a low population density of each taxon, (4) low structuring in the level and the mode of feeding (68% – Sa, 32% – Sb), (5) considerable predominance of small-shelled forms, and (6) lack of manifestations of symbiotic relationships and any evidence of utilization of detritus in food chains. The taxa that constitute this community were mainly adapted to unstable environmental conditions, such as mobile ground; varying salinity and hydrodynamics; and, apparently, a prevalence of poorly transparent water.

2. Developing community. Intermediate stage. The given community was reconstructed on the basis of the oryctocoenosis in Bed 7 (Fig. 3b). It includes 10 taxa,

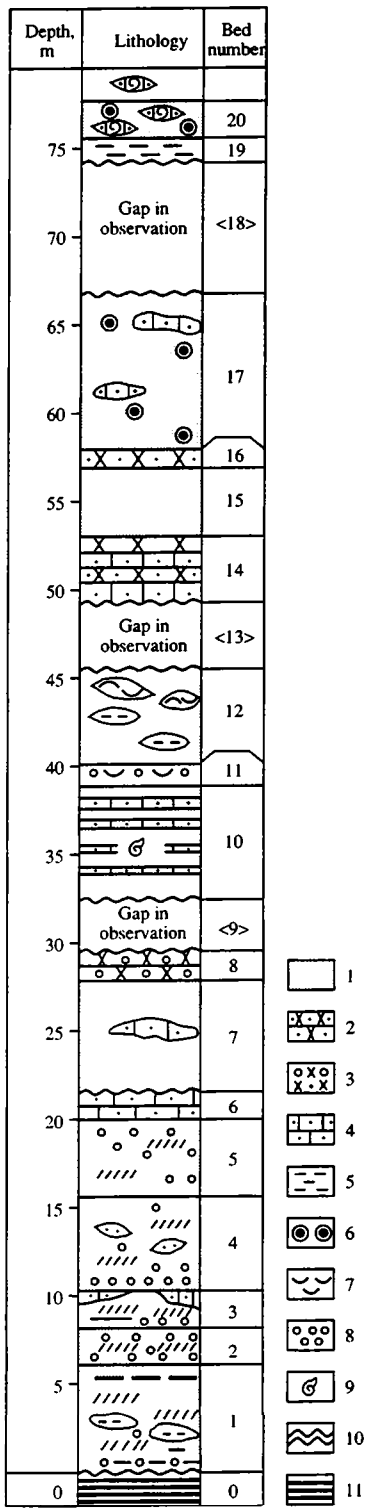

Fig. 2. Composite stratigraphic column of the *Dorsoplanites maximus* Zone (the middle Volgian Substage) in the northern Taimyr Peninsula.

1 – Sand; 2 – sandstone; 3 – gravely sandstone; 4 – sandy limestone; 5 – silt; 6 – sandstone concretions; 7 – coquina composed mainly of oyster shells; 8 – pebble; 9 – occurrences of ammonites; 10 – gaps in observations; 11 – black Silurian shale. Bed numbers are after Mesezhnikov (1984), with additions by the author.

which are suspension feeders; three bivalve genera are suspension feeders of the lower level (*Astarte*, *Pleuromya*, and *Arctica*) and the others represent filter feeders of the upper level (*Ostrea*, *Buchia*, terebratulid *Uralorhinchia*, *Striomodiolus*, *Isognomon*, *Mclearnia*, and *Boreioxytoma*). Despite the higher taxonomic diversity, the appearance of forms with large shells (*Mclearnia* and *Isognomon*), and the somewhat more complex structure in comparison with the pioneer community (for example, the considerably less levelled species ratios and the higher number of trophic levels), elements of the early stage of succession are evident in the developing community. Thus, the role of those groups that first settled in the new biotopes (*Astarte*, *Ostrea*), increased. It is these taxa that form the core of the developing community. Two other pioneer taxa, *Buchia* and *Uralorhinchia*, retained their position. The newly appeared vagile suspension feeders of the lower level (Sa) (*Pleuromya* and *Arctica*) become typical elements of the forming community, whereas new suspension feeders of the higher level (Sb) (predominantly sessile forms) still play an insignificant role, although they entered ecological niches that were new with regard to trophic level, ground type, and possibly water salinity.

The community that temporally follows the previous one may be regarded as paleobiocoenosis only with major reservations, because its reconstruction was based on the coquinas of Bed 10 (Fig. 3c). As this oryctocoenosis is dominated by allochthonous elements, there is no proof that all the species that are shown in the graph lived together under natural conditions, particularly those of rare occurrence.

Nevertheless, this association presents a number of specific characters, like a further increase in taxonomic diversity and increasing structuring in the trophic levels (by the capture of intermediate levels among Sb) and feeding adaptations (appearance of phytofagans represented by *Hudlestonella* and caplike gastropodes). These features make it possible to consider the described benthic community as intermediate between the previous one and the mature community of soft fine-grained sandy grounds, which is considered below.

3. The mature stage. The reconstruction of this community was based on the oryctocoenosis from Bed 17. It includes about 20 taxa of the generic level (Fig. 3d). Though all invertebrates are regarded as detritovores and filter feeders, it is safe to suggest that the numerous and diverse brachiopods (articulate and inarticulate) and lugworms had a diet and, to be sure, mode of food capture that was different from those of bivalves.

Thus, a considerable structuring of the benthic community in both food type and modes of food capture is evident. The community is completely dominated by bivalves of the *Arctotis* species with a moderately large shell. Earlier communities were dominated by small-shelled *Astarte* together with common *Mclearnia*,

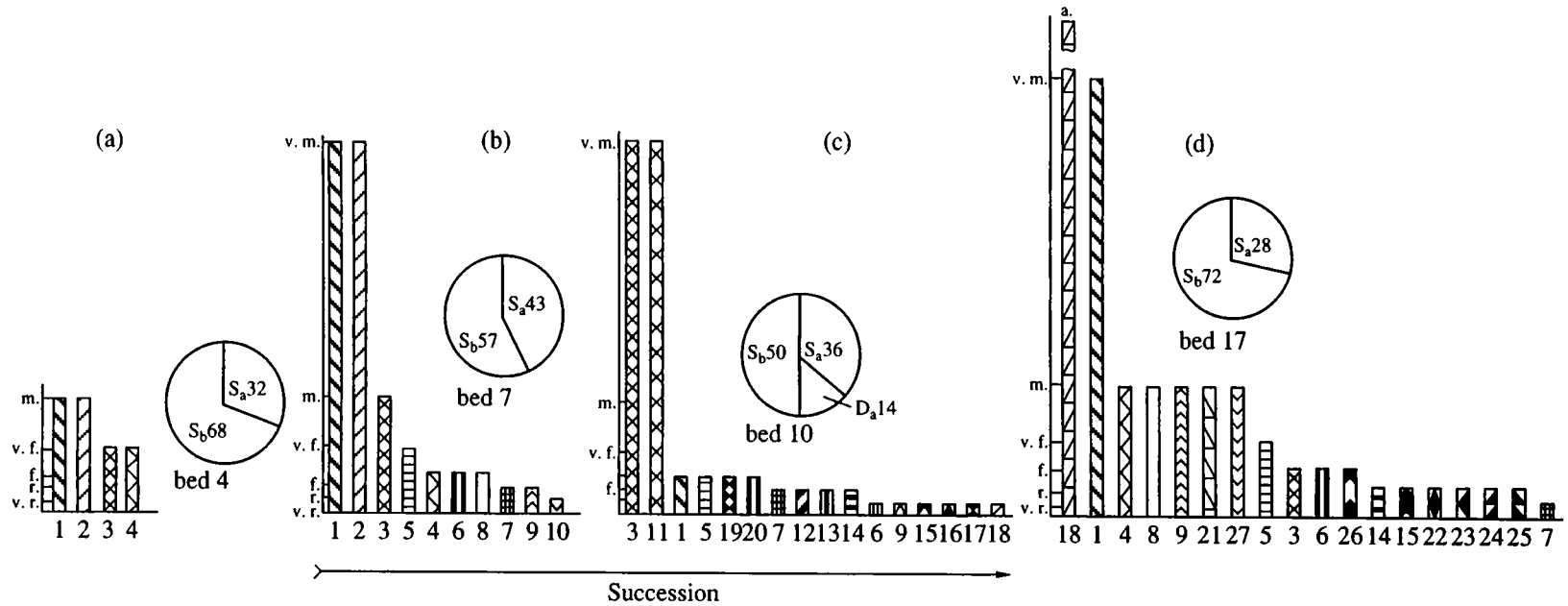

Fig. 3. Succession of benthic assemblages in the Late Jurassic shallow marine bay in the northern Taimyr region.

Histograms: v. r. – very rare (1), r. – rare (2), f. – frequent (3), v. f. – very frequent (5), m. – many (9), v. m. – very many (10), a. – abundant (100). All the taxa, except the specified ones, represent bivalves: 1 – *Astarte*, 2 – oyster, 3 – *Buchia*, 4 – *Uralorinchia* (brachiopod), 5 – *Pleuromya*, 6 – *Striomodiolus*, 7 – *Isognomon*, 8 – *Arctica*, 9 – *Mclearnia*, 10 – *Boreioxytoma*, 11 – *Lopatinia*, 12 - 13 – Chlamys species, 14 – limidae (*Limea* and *Pseudolimea*), 15 – *Aguilerella*, 16 – *Goniomya*, 17 – *Oxytroma*, 18 – *Arctotis*; 19 – (?) caplike gastropod, 20 – *Hudlestonella* (gastropod), 21 – *Entolium*, 22 – *Modiolus*, 23 – *Cosmetodon*; 24 - 25 – two rhynchonellid genera (brachiopods). Circle diagrams: S_a – suspension feeders of the lower level, S_b – suspension feeders of the upper level, D_a – detritovores of lower level; numbers show percentage of a specific group in communities.

Entolium, and *Striomodiolus* with large shells. It is worth noting that, contrary to the previous communities, both the lower and upper level suspension feeders are dominated by vagile taxa: *Astarte*, *Arctica*, *Mclearnia*, *Entolium*, and worms (? lugworms). The very low values of levelling of species ratios are also a good indication of the mature community. The indirect evidence for the increasing pressure of *K*-selection at the formation stage of the community is the decreased role of buchias. Studies of many hundreds of Late Jurassic and Early Cretaceous communities show that these bivalves normally dominate under the leading role of *r*-selection in marginal facies with a marked disbalance of one of the environmental factors (Zakharov, 1981).

DISCUSSION

The four above-described benthic invertebrate communities, which replaced each other in a relatively short (geologically speaking) period, may be considered as a primary ecological succession. Its development may be described by at least three features that are used in the description of modern ecosystems (Odum, 1959, p. 239).

The development of ecosystems is a regular process, manifesting itself in time-dependant changes of species structure and other changes proceeding in communities. The processes are directed and, thus, can be predicted.

Almost all of the mentioned features of the processes are observable in the series of four benthic communities of the middle Volgian Bay in the northern Taimyr Peninsula. These are the following: (1) a directed increase of taxonomic diversity from the four taxa in the pioneer community (I) to 18 taxa in the mature one (IV); (2) the levelling of species ratios decreases; (3) the growing structuring of the community proceed parallel to growing specialization within niches, namely, an increasing number of trophic levels; and (4) the appearance of more diverse modes of food capture. The food chains become more complex with the increasing role of large-shelled invertebrates, the growing role of detritus in the food chains, the appearance of symbiotic forms, and the increasing role of *K*-selection.

Succession proceeds as a result of changes in the abiotic environment by the community, that is, the succession is controlled by the community, despite the fact that the abiotic environment defines the type of succession, the speed of change, and often sets the limits of its development.

This concept is most vulnerable with regard to successions of marine benthic communities. At first glance, it is even insurmountable. Let us begin the discussion from the second part of the postulate regarding the role of the abiotic environment. The evolution of the marine bay in the northern Taimyr Peninsula shows directed changes in the main environmental characteristics. These changes include the stabilization of ground types: mobile hard grounds with variably sized particles gradually transform into stable soft fine-grained and muddy grounds in the course of the working out and levelling of the profile of equilibrium of the bottom. Highly energetic environments move to the coastal zone, bottom hydrodynamics become quiet, salinity approaches the normal marine values, and the water becomes more transparent.

Is it only the transgression that changes the environmental factors in the mentioned directions? Far from it! Numerous vagile bivalves, gastropods, worms, and apparently arthropods (to say nothing about microorganisms and algae that have still not been detected but surely existed), repeatedly disturbing and passing the ground through their intestines, exerted an active (and highly effective) influence on the substrate. They made the water more transparent through continuous filtration. The excretion of metabolites enriched the water in various products, which stimulated the precipitation and development of both larvae and adult specimens of various organisms. Their excrements after burial maintain the development of microbiota and bacterial communities. Thus, even under marine conditions, the organisms produced, as they do at present, a considerable effect on the environment.

The development culminates in a stabilized ecosystem, which has a maximum biomass (or high content of information) per unit of energy flow and a maximum number of symbiotic ties between organisms.

Clearly, it is impossible to calculate the energy flow and matter balance for fossil communities. However, we can discuss some aspects of the biotic relationships between the organisms in paleocoenoses, including symbiotic ties. Hundreds of examples of symbiotic relationships between invertebrates from the geological past are described in the literature, and a number of publications on this subject show no tendency for a decrease (Zakharov, 1988). Commensal relationships are the most reliably established. Marks of commensal sponges and fastened shells of inarticulate brachiopods (*Discinisca*) are recognized in the mature community on large shells of *Mclearnia braeviauris*. Nevertheless, even the most advanced community IV cannot be assigned to the climax stage of the succession, because there is no evidence of such features of general homeostasis as mutualism and predation. At the same time, evidence of commensalism is not undeniable. This is indirectly confirmed by the structure, which still does not reach the climax condition. The latter may be examplified by the synchronous community that inhabited the Khatanga sea in the central Taimyr region and was reconstructed on the basis of the section at the Dyabaka Tari River, the *Maximus* Zone. This community comprises more than 30 genera of invertebrates with dominating large-shelled bivalves *Mc. braeviauris* and subdominant oysters (*Praexogyra praeanabarensis*) and

very large isognomons (*Isognomon volaticum*) (Zakharov, 1966, p. 138). Thus, community IV may be considered only as a stage that precedes the climax one.

CONCLUSION

The described series of four benthic communities sequentially replacing each other in a small marine bay during the *Maximus* Zone (the middle Volgian, the Late Jurassic) is considered to represent a primary ecological succession related to the population of newly formed biotopes. This proceeded either under conditions of the transgressing sea or the gradual levelling of the surrounding Taimyr land after a short marine ingression.

The term "succession" is applied here to communities of the geological past much in the same way as other neontological terms, such as population, species, community (=biocoenosis), that have been assimilated by paleontology. This means only that the inference of a paleosuccession is based on those features of ecological succession that can be reconstructed from traces preserved in the fossil record.

The reconstructed series of four marine benthic communities and the analysis of their development show that both their quantitative and qualitative changes fit those of modern ecological successions. Thus, the paleosuccessions may serve as models for predicting the historical destination of recent successions caused by both natural catastrophes and human activity.

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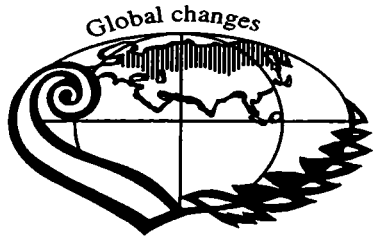

Evolution of the biosphere

Subdivision of Continental Neogene Deposits of the Far East Based on Diatoms

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Abstract – Biostratigraphic subdivisions, such as zones and floral beds, were distinguished in the continental Neogene sequence of the Far East on the basis of freshwater diatom flora. Based on these groups of algae, a zonal scale for the continental Miocene deposits was suggested for the first time. The established subdivisions were used to correlate the main sections of the continental Neogene of the Far East and adjacent regions.

Key words: **Neogene deposits, diatom algae, zonal scale, correlation of sections, the Far East.**

Neogene deposits are widespread in the Far East. They form thick beds in the continental part of the region, adjacent shelf areas, the Kamchatka Peninsula, and the island of Sakhalin. Various mineral deposits are related to these sediments, which stimulated their comprehensive geological study and an elaboration of detailed stratigraphic schemes.

During the last decade, significant progress has been made in studying the stratigraphy of the Far East Neogene deposits of marine origin. In order to subdivide and correlate them, we used zonal scales based on various groups of fauna and flora; diatoms were the most important among them (*Zonal'naya* ..., 1991). In terms of the continental Neogene deposits, the situation is somewhat different. They cover nearly the whole continental area of the Far East and a significant part of the Kamchatka territory, filling large depressions of numerous plain and intermontane basins. However, the stratigraphic scale for these deposits is less detailed than that for the marine beds. This circumstance significantly hinders the subdivision of the continental beds, most of which are coal-bearing. Study of the paleontological remnants from continental Neogene deposits is far from being complete (*Resheniya* ..., 1982a), and this is one of the main reasons for the slow progress in their stratigraphy. First of all, this concerns paleofloral groups, because continental beds in the Far East are extremely poor in fauna. Paleoflora provides the most important criteria for the subdivision and correlation of these deposits, and the majority of their local and regional units are based on the study of macroflora, spores, pollen, and diatoms. Unfortunately, the paleofloral assemblages from the continental Neogene are described only to supplement the paleontological characteristics of one or another previously established unit. So far, there have been few attempts to distinguish independent biostratigraphical units on the basis of the floral groups in order to provide details for stratigraphic schemes of the continental Neogene deposits in the Far East and other regions.

One of the recent attempts concerns the freshwater diatoms. Their long-term use for stratigraphical purposes, which have been described in a series of publications (Moiseeva, 1971; *Diatomovye* ..., 1974; *Stratigrafiya* ..., 1986 and others) as well as their wide distribution and relatively frequent occurrence in the Neogene continental beds offer the possibility of distinguishing supplementary biostratigraphic units, the floral beds (Moiseeva, 1990; Moiseeva and Tsar'ko, 1990), which are traceable throughout the Far East region. Nine beds of this kind were distinguished in the Neogene succession of diatom assemblages in the region. Tracing these beds in the main Neogene sections (the figure), I recognized them in southwestern Primor'e, northeastern Sikhote-Alin', and the southern part of the Khabarovsk region in many local and regional stratigraphic subdivisions.

Further specification of the stratigraphic boundaries of the beds and the intervals of spatial and temporal distribution of characteristic diatom species as well as respective correlations of the main Neogene sections in the Far East (the table) indicate that the succession of diatom-bearing beds and the stability of the composition

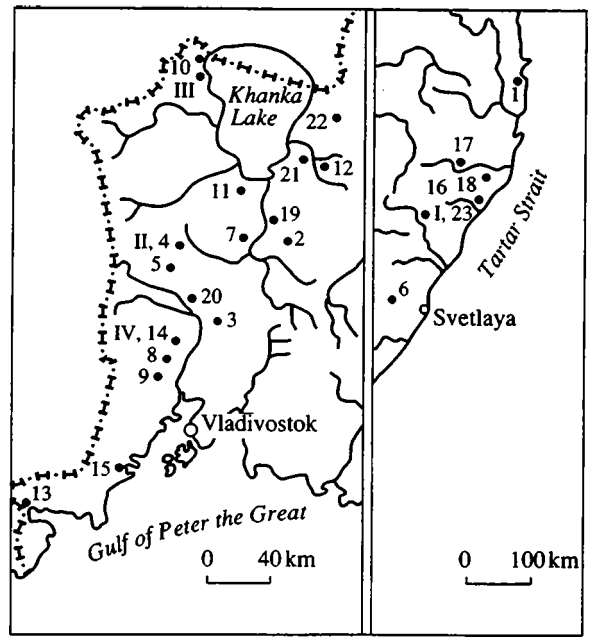
and structure of their diatom assemblages are undoubtedly similar in the Neogene deposits of different areas. At the same time, certain sections with direct stratigraphic contacts of two or more beds can be used to substantiate the position of their stratigraphic limits and the entire succession of the beds from the base upward in the Neogene section.

All of the regularities in diatom distribution provided an opportunity to distinguish units with the main features of biostratigraphic zones (*Stratigraficheskii* ..., 1992) and to elaborate the diatom zonation for the continental Neogene.

A biostratigraphic zone based on freshwater diatoms represents a stratigraphic interval of rocks bearing a certain diatom assemblage with a stable taxonomic composition and structure (Moiseeva, 1993). Species with a wide geographic distribution and relatively short stratigraphic range were included in the zonal assemblages. Such characteristic zonal species are either absent in zonal assemblages of underlying and overlying zones or display there different quantitative and qualitative relations. Only 50 species of this kind were found among a great number of diatom forms known in the continental Neogene. In many cases, infraspecific taxa, the varieties and forms, were attributed to stratigraphic indicators. On the contrary, there are only a few diatom genera that are typical of the Neogene interval, and the generic composition of the Neogene zonal assemblages is usually rather stable. It should also be noted that, not only typical planktonic species, but littoral and benthic forms were also included in the zonal assemblages. They favor the recognition and correlation of diatom zones in continental deposits of both deep-water and shallow-water origin that are spread over vast territories.

The composition of the flora from the diatom zones of the continental Neogene deposits suggests them to be complex zones (coenozones), and, according to paleobiogeographical criterion, they are of the provincial type. Their stratigraphic boundaries correspond to levels of significant changes in the taxonomic composition and structure of the zonal assemblages, resulting from both the evolution peculiarities of this group of algae and the Neogene paleoecological changes. The age of the zones is determined by correlating the zonal assemblages through the main Neogene sections of eastern Russia with due regard for macroflora and palynological data.

The southern part of the continental Far East was suggested to be the main polygon for distinguishing and substantiating the Neogene freshwater diatom zones, because many continental Neogene sections of this territory are well-studied and contain abundant diatom remnants. In order to study the distribution of these zones, the position of their stratigraphical boundaries, and the possibilities of correlating them, I used materials concerning not only the southern Far East, but also adjacent regions: the Baikal and Transbaikal regions



Location map of the zonal stratotypes (zones I - IV) and main sections of diatom beds (1 - 23) in the Neogene deposits of the southern Far East.

I – The Samarga River basin, Toni Lake area; II – Lipovetskaya depression; III – western coast of Khanka Lake; IV – right bank of the Razdol'naya River near Kedrovka Village; 1 – Northeastern Sikhote-Alin', the Aukan Bay area; 2 – Rettikhovskaya depression; 3 – Rakovskaya depression; 4 – Lipovetskaya depression; 5 – Galenkovskaya depression; 6 – Northeastern Sikhote-Alin', the Zeva River basin; 7 – Pavlovskaya depression, 8 – Kedrovka Village; 9 – Nezhino settlement; 10 – western coast of Khanka Lake, near Turii Rog settlement; 11 – Novodevitsa settlement; 12 – Khvalynka settlement; 13 – the Narva River basin; 14 – the Razdol'naya River basin; 15 – the Slavyanka River basin; 16 – the Botchi River basin; 17 – the Koppi River basin; 18 – basins of the Topta and Nemi rivers; 19 – the Ilistaya River basin; 20 – the Mikhailovskaya River basin; 21 – Spasskoe settlement; 22 – Chkalovskoe settlement; 23 – the Toni Lake area.

Names of the zones and beds are given in the table under corresponding numbers.

and the northeastern part of Russia and Kamchatka. Within the studied area, I concentrated on those sections in which diatoms had been studied rather completely and were reliably correlated with local and regional stratigraphic scales. Cheremisinova's and Chernyaeva's data on the Neogene diatoms (Cheremisinova, 1973; *Diatomovye* ..., 1974; *Stratigrafiya* ..., 1986; *Istoriya* ..., 1988; *Paleolimnologicheskie* ..., 1989; *Resheniya* ..., 1981) were used for the Baikal and Transbaikal regions. In addition, I used the local stratigraphic scheme of the Neogene deposits of the Chara depression suggested by Enikeev.[1] I also used Nevretdinova's data on

[1] The scheme was adopted from "Regional Stratigraphic Scheme of the Transbaikal Neogene Deposits," approved by the plenary session of the Interdepartmental Stratigraphic Committee (ISC), November, 1991.

Correlation scheme for stratigraphic units of the continental Neogene in the Far East and adjacent territories based on diatoms

System	Series	Subseries	Provincial diatom zones and beds		Far East						
					horizons	Southwestern Primor'e		Northeastern Sikhote-Alin'		Middle-Amur depression	
Neogene	Pliocene	upper			Suifun	Suifun Fm.		Sovetskaya Gavan' Fm.		Priamur'e Fm.	
		lower		*A. italica–Tet. lacustris* v. *fossilis* Beds (19, 20, 21, 22)			*A. italica–Tet. lacustris–Fr. bicapitata* v. *fossilis* Beds	Andakha Fm.	*A. italica–Tet. lacustris* Beds		
	Miocene	upper	*Actinocyclus gorbunovii* Zone (IV)	*Act. gorbunovii–Tet. celatom* Beds (14, 15, 16, 17, 18)	Ust'-Suifun	Ust'-Suifun Fm.	*Act. gorbunovii–Tet. celatom–G. lingulatum* Beds	Botchi Fm.	*Act. gorbunovii–Act.* sp. Beds	Golovino Fm.	*Act. gorbunovii–Tet. celatom* Beds
		middle	*Ellrbeckia kochii* Zone (III)	*A. praegranulata* v. *praeislandica–Fr. miocenica* Beds (10, 11, 12, 13)	*Fagus chancaica* Beds	Tuffogenic-diatomite deposits	*A. praegranulata* v. *praeislandica–Fr. miocenica* Beds	Maksimovka (upper part) and Kizi Fms		Ushumun Fm.	*A. praegranulata* v. *praeislandica* Beds
				Ellerbeckia kochii–Ach. aengimatosus Beds (10)			*Ell. kochii–Ach. aengimatosus* Beds				*Ell. kochii–Ach. aengimatosus* Beds
			Alveolophora areolata Zone (II)	*Al. areolata–Fr. triandulata* Beds (4, 5, 7, 8, 9)	Khanka region	Upper Pavlovka Subfm.	*Al. areolata–Fr. triangulata* Beds				
				A. semilaevis–Tab. celatom Beds (4, 5, 6)		tuffogenic-sedimentary deposits	*A. semilaevis–Tab. celatom* Beds				
			Alveolophora jouseana Zone (I)	*A. praegranulata* v. *praeangustissima–E. obsoleta–Tab. tumidula* Beds (2,3)		tuffogenic deposits with flora	*A. graegranulata* v. *praeangustissima–E. obsoleta–Tab. tumidula* Beds				
				Al. jouseana Beds (23)					*Al. jouseana* Beds		
		lower		*A. canadensis–Al. bifaria* Beds (1)		Tuffogenic-sedimentary deposits			*A. canadensis* Beds		

(Contd.)

Baikal and Transbaikal regions					
group of Tunkin–Baikal depressions and Ol'khon region		Vitim Upland		Chara depression	
Anosovka and Akhalik Fms, Sasino Member	*A. baicalensis–Cyclotella baicalensis–Stephanodiscus* assemblage	Kydzhimit Member and China Fm.	*A. baicalensis–Stephanodiscus* assemblage	Lyuksugun Fm.	*A. baicalensis–Sibiriodiscus varians* assemblage
			A. italica–Cyclostephanos costata–Tet. lacustris Bed		
Tankhoi Fm., Tagai Member	*Act. gorbunovii–A. praegranulata–Tet. japonicus* assemblage	Dzhilinda Fm.	*Act. gorbunovii–Act* sp. Beds	Anarga Fm.	*Act. gorbunovii–Act.* sp.–*Tet. ellipticus* v. *rostelatus* Beds
	Ell. kochii–Melosira undulata assemblage		*Ell. kochii–A. praegranulata* v. *praeislandica* Beds		*Ell. kochii–A.* sp.–*Tet. lacustris* v. *compactus* Beds
			Al. areolata–A. baicalensis Beds		
			Al. jouseana Beds		

(Contd.)

Northeastern Russia											
horizons	Northern Sea of Okhotsk region (Gizhiginskaya, Penzhinskaya and Nagaevskaya bays)		the Kolyma River basin, Malyk-Sienskaya depression, the well Nizhnekolymskaya-1				Chukotka (Gulf of Anadyr, basins of the Vankarem and Keskuveem rivers)			
Begunovskaya	Sandy-clayey deposits, the Yama River	*A. praegranulata* v. *praeislandica–Tet. lacustris–E. majuscula–Stephanodiscus* assemblag	Kutuyakh Fm.				Sandy-silty deposits		*A. praegranulata* v. *praeislandica–Tet. lacystris–E. majuscula–Stephanodiscus* assemblage	
			Begunovskaya Fm. (core I, 1.60 - 44 m), pebble deposits, M.-S. depression		*A. italica–Tet. locustris–Tet. emarginatus* Beds					
Khapchan			Pokhod Fm. (well I, 120 - 70 m)							
Mamontova Gora	Impoveem and Nagaev-skaya Members, black clays of the Maimakan River	*A. praegranulata* v. *praeislandica–Fr. miocenica* Beds	Stadukha Fm.	Well I, 152 - 135 m	*A. praegranulata* v. *praeislandica–Ell. kochii–Ach. aengimatosus* Beds					
		Ell. kochii–Ach. aengimatosus Beds								
Il'dikilyakh	Shestakovka Member	Assemblage of marine diatoms		Well I, 204 - 176 m	*A. semilaevis–Tab. celatom* Beds					
				Well I, 296 - 228 m	*A. graegranulata–E. obsoleta–Tab. tumidula* Beds		Volch'ya deposits		*A. jouseana* Beds	
		Al. bifaria–A. canadensis Beds								

(Contd.)

Kamchatka								
horizons	Western Kamchatka (Tochilinskii section)		Southeastern and Central Kamchatka (Kronotskii Gulf, Central Kamchatka depression, and Kamchatka River basin)		horizons	Northeastern Kamchatka (Bay of Korf)	
Enemten	Enemten Fm.	Assemblage of marine diatoms of the *D. kamtschatica* Zone	Storozhka Fm. / Shchapinsk Fm.	*A. italica–Tet. lacustris–Cyclostephanos costata–Stephanodiscus–Cyclotella* assemblage	Limimtevayam and other		
Erman	Erman Fm.	*A. praegranulata–Act. gorbunovii–Undatodiscus variabilis–Tet. celatom* assemblage			Klassicheskii	Upper Medvezhka Subfm. and Klassicheskii Fm.	*A. praegranulata–Act. gorbunovii–Undatodiscus variabilis* assemblage
Etolona		Marine diatom assemblage of the *D. hustedtii - D. lauta* Zone	Volcanogenic-sedimentary deposits of the Kamchatka River	*A. praegranulata–Ell. kochii–Tet. ellipticus v. lancea–Tet. lacustris v. lanceolata* assemblage	Medvezhka	Lower Medvezhka Subfm.	–
Kakertok		Marine diatom assemblage of the *D. lauta* Zone	Mokovo Fm.		Pakhacha	Uspenskoe Fm.	Marine diatom assemblage of the *Stephanopyxis schenckii–Goniothecium tenue* Zone
Kuluven		Marine diatom assemblage of the *Paralia sulcata–Kisseleviella cariina* Zone					

The adopted abbreviations: *A – Aulacosira; Ach. – Achnanthes; Act. – Actinocyclus; Al. – Alveolophora; D. – Denticulopsis; E. – Eunotia; Ell. – Ellerbeckia; Fr. – Fragilaria; G. – Gomphonema; Gon. – Goniotecium; St. – Stephanopyxi; Tab. – Tabellaria; Tet. – Tetracyclus;* Fm. – formation; Subfm. – subformation.

the Neogene diatoms of northeastern Russia (Nevretdinova, 1982, 1986; Sadykov and Nevretdinova, 1984, 1988; Bespalyi *et al.*, 1984; *Istoriya ...*, 1988; *Paleogen ...*, 1989; Volobueva *et al.*, 1990) and Lupikina's determinations of the Miocene flora from the Nagaevskaya Bay (Chelebaeva *et al.*, 1979). Finally, my knowledge of the Kamchatka area is based on Lupikina's and Ozornina's data (Shantser *et al.*, 1969; *Diatomovye ...*, 1974; Sinel'nikova *et al.*, 1976; Lupikina, 1980; *Resheniya ...*, 1982b; *Atlas ...*, 1984; Ozornina, 1987)

A description of the diatom biostratigraphic zones from the base upward in the Neogene sequence is given below.

Alveolophora jouseana Zone (end of the early – beginning of the middle Miocene).

This zone is characterized by a diatom assemblage dominated by the zonal index-species, which has a wide interregional distribution in the continental deposits of the early-middle Miocene. Extinct Neogene species *Aulacosira praegranulata* (Jousé) Simon and especially its type variety *Tetracyclus ellipticus* (Ehr.) Grun. var. *ellipticus* et var. *linearis* Hust. are the subdominant forms of the assemblage. A few specimens of *Eunotia japonica* Pant., *E. polyglyphoides* Sheshuk., and *Gomphocymbella miocenica* Jousé appear at this stratigraphic level for the first time. These forms are the common accompanying species in the zonal assemblage. In some regions (southwestern Primor'e and the lower Kolyma depression), the index-species is usually absent in the upper part of the zone, where characteristic species are represented by the above representatives of *Aulacosira* and *Tetracyclus* genera and by rarely occurring *Fragilaria constricta* var. *antiqua* Moiss., *F. nitzschioides* var. *kamchatica* Lupik., *Tabellaria tumidula* Moiss., and *Eunotia clevei* var. *sinica* Skv. This is mainly caused by the peculiar regional paleoecology during the accumulation of the uppermost deposits of the zone (Moiseeva and Tsar'ko, 1990).

The stratotype of the zone is represented by a section of the lower part of the Kizya Formation (tufogenic-sedimentary and sedimentary deposits), situated in the northeastern Sikhote-Alin' along the left bank of the Samarga River near Toni Lake (Plakhotnik, 1962).

Outside the Sikhote-Alin', the zone is traceable in Primor'e within the tuffogenic flora-bearing sediments of the Rettikhovskoe brown coal deposit and the tuffogenic barren unit of the Rakovskaya depression. The lower part of the Dzhilinda Formation, represented by alternating claystones, siltstones, and diatomites that fill some depressions of the Vitim Highland, is an analogue of this zone in the Transbaikal region. In northeastern Russia, it has been traced within the sandy-clayey deposits of the Volch'ya unit of the Gulf of Anadyr coast and in the intercalating clayey, silty, and fine-sandy beds with plant detritus found in the Kolyma River basin. The latter were penetrated by the well Nizhnekolymskaya-1 at a depth of 228 - 226 m and, most likely, correspond to the Stadukha Formation. The *Al. jouseana* Zone is absent in Kamchatka, because diatom-bearing beds of this stratigraphic level are of marine origin there.

The taxonomic composition of the zonal assemblage and its position within the continental Neogene sequence of the Far East (the table) suggest that the age of the *Al. jouseana* Zone ranges from the second half of the early to the beginning of the middle Miocene. In general, this age determination correlates with data on abundant macroflora and palynological estimates. The only exception is the Kolyma basin, where the spore and pollen assemblage from the described interval of the drilled section is dated at the end of the late Oligocene – beginning of the early Miocene (Fradkina *et al.*, 1989).

Alveolophora arealata Zone (middle Miocene).

The extinct Miocene species *Alveolophora areolata* (Moiss.) Moiss. is the most typical species of the zone. Being genetically close to the index-species of the *Al. jouseana* Zone, it is distinguished by a more complex morphology of the test. Sporadically occurring in the underlying zone, species *Aulacosira semilaevis* (Grun.) Moiss., *Melosira undulata* var. *samoensis* Grun., *Tabellaria celatom* Churs., and *Eunotia monodon* var. *tropica* Hust. predominate in this assemblage. Several varieties of *Aulacosira praegranulata* (Jousé) Moiss. remain subdominant. Species with a narrow stratigraphic range, such as *Fragilaria triangulata* Moiss. and *Navicula bituminosa* Pant. var. *bituminosa* et var. *latecapitata* Pant., are characteristic only for this zone. In addition, the first appearance of species that predominate in the assemblages of higher stratigraphic levels is recognized here. They are represented by *Ellerbeckia kochii* (Pant.) Moiss., *Fragilaria bicapitata* var. *fossilis* Moiss., *F. miocenica* Jousé, *Tetracyclus celatom* Okuno, *Actinella brasiliensis* var. *cuneata* Moiss., *Achnanthes aengimatosus* Moiss., and *Navicula amphibola* Cl.

A more thermophylic character distinguishes the described zonal forms from the other Neogene diatom assemblages. It is manifested in the complete absence of species of the cold-water *Tetracyclus* genus in the lower part of the zone and in their low content in the upper part. Such thermophylic varieties and forms as *Melosira undulata* (Ehr.) Kütz. var. *undulata* et var. *samoensis* Grun., *Actinella brasiliensis* Grun. var. *brasiliensis* et var. *cuneata* Moiss., *Eunotia clevei* var. *silica* Skv., *E. monodon* var. *tropica* Hust., and *Navicula bituminosa* Pant. var. *bituminosa* et var. *latecapitata* Pant. are important elements of the zonal assemblage. The described peculiarity was probably caused by a relative climatic warming (most likely by the climatic optimum) during the accumulation of the assemblage.

The section of the tuffogenic-sedimentary member of the upper Pavlovka Subformation described by Baskakova and Gromova (1989) at drilling site 1458 of the Lipovetskaya depression in Primor'e is the stratotype of the zone. The lower zonal boundary is distin-

guishable here. The contact between the lower part of the zone, which is represented by the *Aulacosira semilaevis–Tabellaria celatom* Beds, and the underlying *Al. jouseana* Zone is also traceable in the Nizhnekolymskaya-1 well section at a depth of 204 m.

In addition to its frequent occurrence in southwestern Primor'e, this zone was also recognized in many sections of the sedimentary-volcanogenic Dzhilinda Formation of the Vitim Upland, the Transbaikal region. The presence of *Aulacosira baicalensis* (Wisl.) Simon., the present endemic of Lake Baikal, is a specific feature of the zonal assemblage. This zone seems to be the first appearance level of this species. In northeastern Russia, the lower part of the *Al. areolata* Zone was traced in a member of light to dark grey clays of the Stadukha Formation, at a depth interval of 204 - 176 m at the drill site Nizhnekolymskaya-1. In Kamchatka, the diatom-bearing continental Neogene deposits of this level have not yet been found.

The taxonomic composition of the described zonal assemblage and the stratigraphic ranges of its main representatives in the local and regional Neogene units of the Far East and Transbaikal regions suggest that the zone is the middle Miocene in age.

Ellerbeckia kochii Zone (late middle Miocene).

This zone and the underlying ones are nearly identical in their species composition. However, the relative role of certain species in the structure of the new zonal assemblage is different. Extinct species *Ellerbeckia kochii* (Pant.) Moiss., *Melosira undulata* var. *samoensis* Grun., and *Achnanthes aengimatosum* Moiss. are the most characteristic here. Varieties of the extinct planktonic species *Aulacosira praegranulata* (Jousé) Simon. and especially var. *praeislandica* (Jousé) Moiss. become dominant in the upper part of the zone. Subdominants include varieties and forms of *Tetracyclus ellipticus* (Ehr.) Grun., mainly var. *lancea* (Ehr.) Hust. and extinct variety *T. lacustris* var. *lanceolatus* Moiss. During the evolution of the *Tetracyclus* genus, the latter was a linking form between the extinct Neogene species *T. ellipticus* (Ehr.) Grun. and *T. lacustris* Ralfs, the present dweller of oligotrophic lakes. The extinct *Fragilaria miocenica* Jousé also belongs to the subdominant species. At this stratigraphic level, it reaches its maximum abundance. Such species and varieties as *Ellerbeckia teres* (Brun) Crawf., *E. arenaria* (Moore) Crawf., *Aulacosira italica* var. *valida* (Grun.) Simon., and *Tetracyclus lacustris* Ralfs var. *lacustris* et var. *strumosus* (Ehr.) Hust. appear among the accompanying forms of the zonal assemblage for the first time, especially in the upper part of the zone. The flourishing stage of these forms is recorded in the younger Cenozoic beds and in the present diatom flora. In the Far East, the zone corresponds to the first appearance level of the extinct freshwater species of the *Actinocyclus* genus. These species became more abundant in the younger late Miocene flora. It is evident that inheritance occurred between the lower part of the zone and the *Alveolophora areolata* Zone and between its upper part and the upper Miocene *Actinocyclus gorbunovii* Zone, which provides characteristic features to the taxonomic composition of the zonal assemblage.

The stratotype of the zone is represented by the section of the tuffogenic-diatomite unit described by Baskakova and Gromova (1986) in Primor'e on the western coast of Khanka Lake between Novokachalinsk and Turii Rog settlements. Stratigraphically, this unit belongs to the regional *Fagus chancaica* beds (*Resheniya* ..., 1982a).

The following layers are exposed in this section (from the base upward):

(1) Volcanomictic conglobreccia with interlayers of inequigranular tuffaceous sandstone and tuffaceous siltstone. Thickness: 100 m.

(2) Unexposed part of the section. Thickness: 150 m.

(3) Tuffaceous sandstones and siltstones with ashy tuff interbeds. Thickness: 50 m.

(4) Intercalation of brownish-grey, grey, and light grey diatomites, tuffaceous diatomites, tuffaceous siltstones, and tuffites with abundant leaf imprints and macroflora remnants. Thickness: 200 m.

(5) Tuffaceous sandstones and siltstones with ashy tuff interbeds. Thickness: 40 m.

(6) Sandy-pebbly unit with thin interlayers of siltstones, clays, and tuffs. Thickness: 250 m.

The upper boundary of the zone coincides with the top of a 40-m-thick unit of tuffaceous sandstones and siltstones with ashy tuff interbeds. Direct contact of the zone with the underlying *Alveolophora areolata* Zone is observed within the tuffo-diatomite unit of the drilled sections near Novodevitsa settlement and in the section of the Dzhilinda Formation recovered by drilling in the Amalatskoe plateau of the Transbaikal region.

In Primor'e, the zone was traced not only across the western coast of Khanka Lake, but also in the Narva River basin within sandy-siltstone and claystone deposits, which Klimova (1983) suggested naming the Sinii Utes Formation. It was also traced in a tuffogenic-sedimentary member represented by sandstones, siltstones, and diatomites, probably of the same formation, which was recovered by drilling and exposed in the Khanka plain near Khvalynka and Spasskoe settlements. A member of intercalating silty clays, sands, and diatomites of an upper part of the Ushumun Formation corresponds to this zone in the Middle-Amur depression. In the Baikal and Transbaikal regions, a similar zonal assemblage is relatively abundant in the upper part of the Tankhoi Formation and the Tagai member of the Ol'khon area and Tunkin–Baikal group of depressions. It also was recognized in the upper part of the Dzhilinda Formation of Vitim Upland and in the Anarga Formation of the Chara depression. The diatom assemblages from the Nagaevskaya member of the Nagaevskaya Bay area near Magadan (Chelebaeva *et al.*, 1979) and from the black clay member of the

Maimakan River basin of the Gizhiginskaya Bay coast (Nevretdinova, 1982) are characteristic of this zone in northeastern Russia. The diatom assemblage from the Ipoveem member of the Taigonos Peninsula in the northern Sea of Okhotsk area probably corresponds to the lower part of the *Ellerbeckia kochii* Zone, though it has been suggested that it was formed during earlier Miocene epochs (Sadykov and Nevretdinova, 1988; Volobueva *et al.*, 1990). Despite the relatively endemic taxonomic composition, this assemblage is correlative with the *E. kochii* zonal assemblage due to the abundance of such characteristic species as *Ellerbeckia kochii* (Pant.) Moiss., *Achnanthes aengimatosus* Moiss., *Tetracyclus ellipticus* var. *lancea* (Ehr.) Hust, and others. Presently, the age of the Ipoveem assemblage has to be refined by additional and more detailed studies of its taxonomic composition and stratigraphic range. The upper part of the Stadukha Formation corresponds in the Kolyma River basin to the zone under consideration. It is represented by grey and light grey clays with thin interlayers of fine sandstones recovered by the well Nizhnekolymskaya-1 at a depth of 152 - 135 m. In southeastern Kamchatka, a diatom assemblage presumably corresponding to the zone was found in some sections of the volcanogenic-sedimentary deposits in the basin of the Kamchatka River and its tributaries (*Resheniya* ..., 1982a). However, its taxonomic composition and position within the local stratigraphic scheme need to be studied in more detail.

The peculiar composition of the assemblage, its concrete stratigraphic position in many sections, and the reliable paleobotanic substantiation of the middle Miocene age inferred for the *Fagus chancaica* Beds (*Resheniya* ..., 1982a) suggest that the *E. kochii* Zone corresponds to the second half of the middle Miocene.

Actinocyclus gorbunovii Zone (late Miocene).

This zone is characterized by a rather specific and diverse diatom assemblage. Extinct freshwater species of the *Actinocyclus* genus, namely, *A. gorbunovii* (Sheshuk.) Moiss. et Sheshuk., *A. tubulosus* Churs., and *Actinocyclus* sp., predominate in it. These species occur in the top middle Miocene deposits only sporadically. Here, they reach their maximum abundance. The species *A. praegranulata* var. *praeislandica* (Jousé) Moiss. is still the abundant representative of the ancient extinct forms of the *Aulacosira* genus. Subdominant and constantly accompanying forms of the assemblage include the following: *Fragilaria bicapitata* var. *fossilis* Moiss., *Fragilaria lineolata* (Moiss.) Moiss., *Diatoma capitata* Lauby, *Tetracyclus celatom* Okuno, *T. lacustris* var. *strumosus* (Ehr.) Hust., *Eunotia polyglyphoides* Sheshuk., *E. clevei* var. *aculeata* Moiss., *E. veneris* var. *nipponica* Skv., *Achnanthes lapidosa* f. *robusta* Moiss., *A. scutiformis* Moiss., and *Gomphonema lingulatum* Hust. Most of them are the extinct forms or extremely rare representatives of the present flora. The rare extinct species *Gomphopleura poretzkiae* Moiss. was found only at this level. Many species from earlier Miocene levels, such as *Ellerbeckia kochii* (Pant.) Moiss., *Actinocyclus* sp., *Tetracyclus lacustris* var. *lanceolatus* Moiss., and *Fragilaria miocenica* Jousé terminated their evolution within this zone. These species and the extinct form *Aulacosira praegranulata* (Jousé) Simon. are evidence of close relations between this zone and the underlying *Ellerbeckia kochii* zone. The role of the extinct Neogene forms of *Tetracyclus ellipticus* (Ehr.) Grun. decreases in the zone, while that of the presently existing *T. lacustris* species increases. The first appearance of many new species of the *Fragilaria*, *Eunotia*, *Navicula*, *Pinnularia*, and *Gomphonema* genera is confined to this level. They reached their maximum abundance later, mainly in the recent diatom flora.

The stratotype of the zone is represented by the section of diatomites and clays of the Ust'-Suifun Formation in Primor'e, situated on the left bank of the Razdol'naya River, 1.5 km northeastward from the Kedrovka Village, near the Pionerskoe diatomite deposit. The following layers are exposed in the section (from the base upward):

(1) Clay, brown, sometimes ochreous. Thickness: 0.30 m.

(2) Diatomite, plastic, dark grey and slightly creamy colored in the lower part, and light grey with thin interlayers of fine sands in the upper part. Thickness: 6 m.

(3) Sandy clay with thin diatomite interlayers. Thickness: 1 m.

(4) Loams and loamy sands with rock debris, pebbles, and boulders. Thickness: 0.5 m.

(5) Unexposed interval, 0.20 m wide.

The lower boundary of the zone is established at the base of the Ust'-Suifun Formation, which was penetrated by drilling near Nezhino and Razdol'noe settlements. The direct contact of the described zone with the underlying *Ellerbeckia kochii* Zone is also traceable in the upper part of the Dzhilinda Formation, exposed and recovered by drilling in the Amalatskoe plateau of the Transbaikal region.

In southwestern Primor'e, the zone is recognizable only in the section of the Ust'-Suifun Formation, but was distinguished in many sections of the Pushkinskaya depression and in the Razdol'naya and Slavyanka river basins. A similar assemblage with a slightly less diverse taxonomic composition characterizes the deposits of the Botchi Formation in the northeastern Sikhote-Alin' and the Golovino Formation of the Middle Amur depression. In the Baikal region, the zone was traced in many sections of the upper part of the Tankhoi Formation in the Tunkin and Baikal depressions. In the Transbaikal region, sediments from an upper part of the Dzhilinda Formation of the Vitim Upland and those from the Anarga Formation of the Chara depression correspond to this zone. The described zone has not yet been established in northeastern Russia. In Kamchatka, similar assemblages were traced along the western coast of the peninsula in the sediments of the Erman Formation

and in its northeastern part along the coast of the Bay of Korf in the sediments of the Upper Medvezhka Subformation and the Klassicheskii Formation. The zonal index-species predominates in the Kamchatka assemblage along with the closely related species *Undatodiscus variabilis* Lupik. Lupikina suggested that the interval of the Erman Formation that is characterized by this assemblage be named the *Melosira praedistans* Beds (*Atlas* ..., 1984). However, this name is hardly adoptable, because many specialists do not believe this species to be taxonomically substantiated. In addition, the presence of *Undatodiscus variabilis* Lupik. is also a typical feature of the Kamchatka assemblage.

The peculiarities of the taxonomic composition and structure of the assemblage and the stratigraphic ranges of its characteristic forms within the Neogene sections of Russia, Western Europe, China, Japan and the USA indicate the late Miocene age of the *Actinocyclus gorbunovii* Zone.

Thus, the analyzed data on the Neogene freshwater diatom flora show that continental Neogene deposits of the Far East are divisible into four biostratigraphic zones, which cover the time interval from the second half of the early Miocene to the late Miocene. The correlation of the zonal assemblages in relation to the local stratigraphic subdivisions of the Baikal and Transbaikal regions, northeastern Russia, and Kamchatka revealed that the positions of the zonal boundaries are relatively stable and can be traced in distant regions. The spatial and temporal distribution of the zones within the continental Neogene sections of different regions of eastern Russia suggest that these units are favorable for correlation analysis.

By now it seems impossible to distinguish such zones for the lower part of the lower Miocene and for the Pliocene. In the first case, there are only scarce finds of diatom remnants of that age. However, in the second case, it is impossible because of the existence of abundant assemblages with a diverse and mixed taxonomic composition, which makes them too complex for correlation. This feature was mainly caused by intense and diverse relief changes during the Pliocene epoch, which resulted in a high degree of ecological endemism of the diatom assemblages not only in different regions, but often within a limited area. For these reasons, paleontological substantiation and correlation of the local lower Miocene and Pliocene stratigraphic subdivisions are still based on supplementary biostratigraphic units, such as floral beds (Moiseeva and Tsar'ko, 1990). It should be noted that an analysis of the materials concerning the Pliocene diatoms of eastern Russia shows that an elaboration of local diatom zones for the continental Pliocene seems to be promising, especially for the area of the Vitim Upland. In this region, diatoms are widespread throughout the Pliocene sequence, thus providing opportunities for distinguishing their most reliable succession.

Most of the described diatom zones and beds promote more detail in the local and regional stratigraphic scales and a more reliable regional and interregional correlation of their subdivisions. Elaboration and improvement of the suggested diatom zonal scale of the continental Neogene should be focused on its lowermost part and several Pliocene levels. Further investigations may also provide grounds for reclassifying some floral beds as zonal or subzonal units.

Reviewer Yu. B. Gladenkov

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Correlation of Loess-Soil Sections in Northern Bulgaria from Magnetic and Paleomagnetic Data

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Abstract – Three loess-soil sections in northern Bulgaria were magnetically and paleomagnetically studied. Up to six horizons of buried soils and loesses were identified in the exposures, and magnetic data were shown to be useful for correlation of the studied sections (the values of magnetic susceptibility and the Koenigsberger factor). The obtained magnetic and paleomagnetic information allowed the conclusion that the lowermost loess-soil horizons are coeval. The interpretation of paleomagnetic data implied that loess formation in northern Bulgaria started in the Eopleistocene and not in the Pleistocene, as is generally accepted.
Key words: **Magnetic susceptibility, magnetization, correlation, stratigraphy, Koenigsberger factor.**

INTRODUCTION

The stratigraphy and correlation of loess-soil sections are universally considered to be important (Veklich, 1982; Velichko *et al.*, 1984; *Stratigrafiya i Paleogeografiya...*, 1992). The paleomagnetic method is practically the only one by which we can locate the chronological boundary of the Pleistocene in loess-soil formation; however, the results of this approach are often controversial (Veklich, 1982; Tret'yak, 1983; Heller and Tungsheng, 1984; Faustov and Virina, 1989). The paleomagnetic studies of loess-soil sections in northern Bulgaria carried out by Wiegank (1990) and Kalcheva *et al.* (1989) may serve as an example. According to their conclusions, the loess complex in Bulgaria relates to the Bruhnes. However, Wiegank identifies five excursions, dating them from 575 ka to 45 ka, whereas Kalcheva and coauthors did not find a single event in a section twice as thick. The Pleistocene age determinations of the loess complex in Bulgaria are substantiated by the geomorphological, faunal, and archeological data of most researchers (Kr'stev and Kr'steva, 1980; Kalcheva *et al.*, 1989; Wiegank, 1990). It is important to note that, according to the results of paleomagnetic research, loess accumulation in nearby regions of central Europe, the south of Eastern Europe, and Central Asia and China (Pevzner and Pecsi, 1980; Pecsi, 1993; Veklich, 1982; Heller and Tungsheng, 1984), started in the Matuyama epoch. Consequently, the paleomagnetic dating of loess accumulation in Bulgaria contradicts the results from other regions of the northern hemisphere.

In our studies of the loess complex of Bulgaria, we endeavored to apply combined magnetic and paleomagnetic data to obtain the correlation and stratigraphy of the loess-soil sections.

MATERIALS AND METHODS

We studied the sections of loess-soil formation in northern Bulgaria exposed in the quarries near the Vetovo and Kaolinovo settlements and on the Black Sea coast near Durankulak. The distances between the sublatitudinally oriented sections are about 70 km and 120 km, respectively. In the Vetovo section (Fig. 1), the deposits of an elevated watershed plain are composed of a series of buried soils (BS) intercalated by loess horizons (L). In all, there are six BS and six L. The two upper chestnut-brown soils are poorly defined. The two middle buried soils of the section belong to a transitional type of soil formation ranging from chestnut to black and red earth. The lower two BS5 and BS6 are distinctly outlined and are mainly red earth. The Kaolinovo section (Fig. 2) exposes deposits of five soil-loess horizons composing a watershed with absolute altitudes of about 270 m. The Vetovo and Kaolinovo sections have many common structural features; they differ in that the former is more stratified and somewhat thicker at the expense of the loess beds. The smaller number of loess-soil horizons in the second section can be attributed to the thinning out to complete disappearance of the loess horizon corresponding to the fifth loess bed in the Vetovo section, and to the imposed effect of the formation of the overlying and underlying soils. This imprint is indirectly implied by the greater thickness of the fourth BS in the Kaolinovo section and by the remains in its middle part of a carbonate (illuvial) horizon. Apparently, this soil is composed of two parts and corresponds to the fourth and fifth BS of the Vetovo section. At the base of both sections, Pliocene yellow-rose colored sandy loam rocks occur. The Durankulak section (Fig. 3) is located in a different geomorphological setting composed of a sea-eroded spur in the loess marine plain on the right bank of the Danube with absolute marks of 10 - 15 m and more. The cliffs in this area reveal the cyclic structure

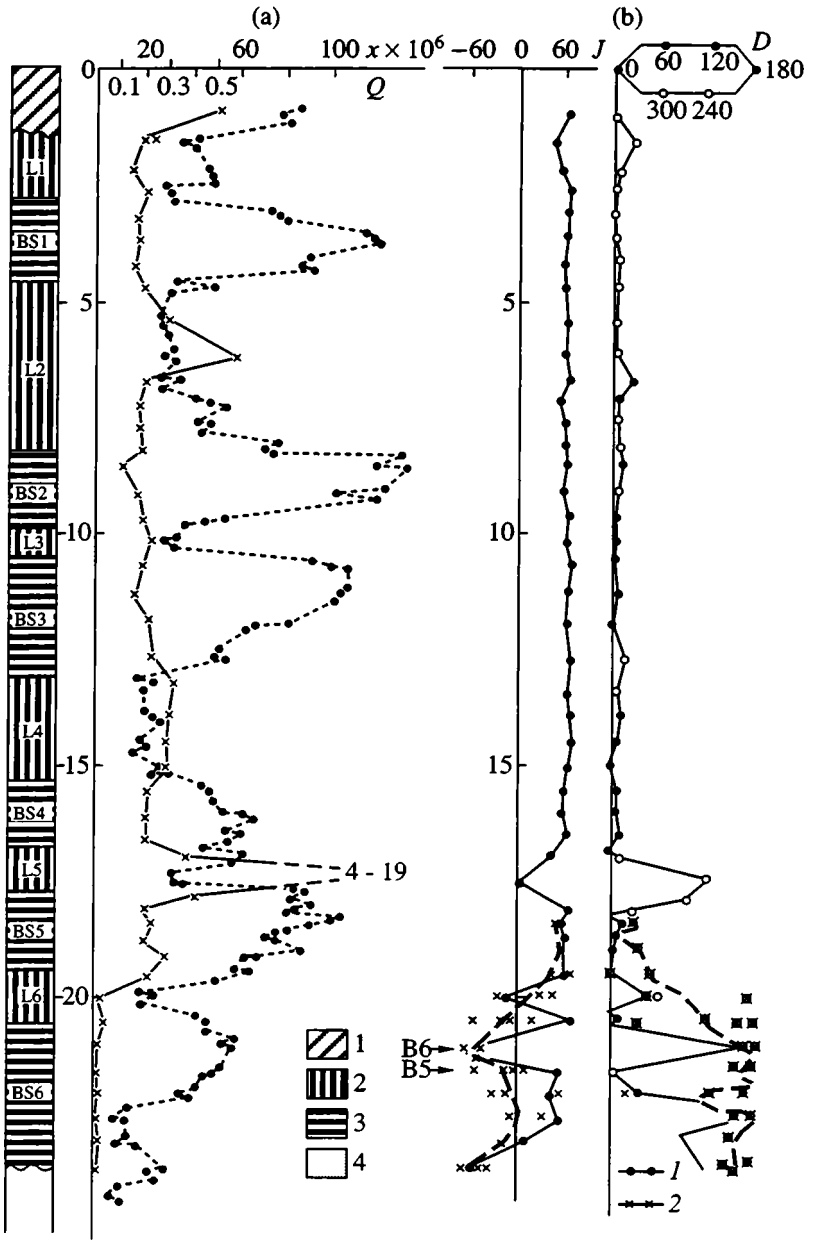

Fig. 1. Lithological, magnetic, and paleomagnetic characteristics of the Vetovo section.
a – Magnetic susceptibility in Gauss units (points) and the Koenigsberger factor after $T = 200°C$ (crosses); 1 – recent soil, 2 – loess, 3 – buried soil, 4 – sandy loam; b – declination and inclination angles after T – demagnetization; *1* – at 200°C, *2* – at 310°C.

of the loess-sandy complex, deposited with deep erosion on the Sarmatian limestones. Every cycle of the section is composed of loesslike sandy loams and loams, which are carbonate and brown and yellow-brown in color, and buried soils. In all, the section has six buried soils of the automorphic type. The two lower red soils are the thickest, and the two in the middle are also well-defined chestnut and brown soils. The upper soils are thin and poorly represented in the section.

The oriented rock samples with dimensions of no less than 10 × 10 × 10 cm^3 were collected at 40- to 60 cm intervals. Between four and 10 cubes with 2.4 cm sides were cut from each sample. The measurement of remanent magnetization was carried out with demagnetizators ION-1 and JR-4; magnetic susceptibility was measured with the IMV-2 and KLY-2 instruments of Soviet and Czechoslovakian serial production, respectively.

RESULTS

The magnetic susceptibility x of the studied sections has higher values in the soils and comparatively lower values in the loesses (Figs. 1 - 3). The value of the

Koenigsberger factor (Q) is also higher in the soils (1.5 - 3) than in the loesses, where it varies from 0.9 to 1.5. Thermomagnetic research indicates the presence of magnetite, maghemite, and hematite in the sediments (Bol'shakov, 1993). Moreover, throughout the section, the dependencies of magnetization saturation on temperature (T) are practically identical for loesses and soils.

The measurement of natural remanent magnetization ($\grave{I}_n$) in all samples, except those collected from the lower part of the fifth loess in the Vetovo section, revealed ($\grave{I}_n$) directions that closely correspond to the direction of the recent magnetic field of the Earth (MFE). A special study (Bol'shakov, 1993) showed that the magnetization anomaly in the samples from the lower part of the fifth loess (angle of declination D = 244° and angle of inclination J = 3°; the value of factor Q is 19) and in those from the upper part of the fifth BS is of "lightning" origin, i.e., the magnetization of the sediments is the result of a thunderstorm discharge (lightning) passing through them.

Temperature-temporal demagnetization was performed by keeping the samples in a nonmagnetic environment at T = 200°C for four hours; after this process, the ($\grave{I}_n$) directions in most of the samples changed only slightly. Angle D deviated irregularly from the initial direction by no more than 15°. In most cases, angle J was reduced by 5° - 10°, but the ($\grave{I}_n$) directions in samples from the lower part of the sections, including sandy loam deposits (the lowermost buried soil) and loess, changed essentially. The sandy loam sediments in the Kaolinovo section (Fig. 2) showed an almost reversed magnetization. The ($\grave{I}_n$) directions in samples from the lowermost loess horizon in all sections changed in both D and J. In the lowermost BS (Figs. 1 - 3), lower values were recorded for the J angle than in the upper part of the section. While D and J exhibit various changes in the lower part of all three sections, the change in factor Q after demagnetization is more consistent. Figures 1 - 3 show an obvious, no less than twofold reduction of Q in samples from the lowermost horizons of buried soil and loess as compared to that of the overlying horizons in all sections. In this way, a definite correlation is indicated between the lowermost loess-soil horizons of the studied sections. This induced us to carry out supplementary temperature measurements of natural remanent magnetization.

Instruments at the Paleomagnetic Laboratory of Kazan' University were used to obtain Zijderfeld diagrams (ZD) for some of the samples. The change in ($\grave{I}_n$) under temperature effect in the samples from the upper part of the section is, in the majority of cases, a virtually straight line in the X–Z plane, which does not cross the coordinate axes and arrives in the area of the beginning of the coordinates. The ZD of the samples taken from the loess and buried soil at the base of the sections have a different aspect. Usually, they show two differently oriented components of magnetization in a sample, one of which (low-temperature) coincides with the direction toward the recent pole. The change to another direction often occurs in the temperature interval 300 - 400°C (Fig. 4). The same procedure was used in a stage-by-stage demagnetization of the samples from the anomalous zone during 30-min intervals at T = 175, 250, and 310°C and subsequent cooling of the samples to room temperature. Higher temperatures damaged the samples, because after T = 310°C, they disintegrated during measurements. The ZD was taken from smaller samples under a continuous temperature rise; this process probably caused their higher resistance to heating. During the stage-by-stage demagnetization, the Q parameter gradually reduced to values of 0.03 - 0.01 and less. In this process, the angles of declination and inclination changed unidirectionally up to T = 310°C. In a strict sense, the results of the stage-by-stage demagnetization do not indicate the achievement of a stable ($\grave{I}_n$) direction after heating to 310°C (the number of points is too small). However, taking into account the ZD data, we can state that heating to that temperature produces ($\grave{I}_n$) directions that are close to stable. On the other hand, maximal heating to 310°C causes the above-mentioned thermal destruction of the samples. Therefore, the overall temperature demagnetization of samples was performed at 310°C for 1.5 hours. The results are shown in Figs. 1 - 3. We should note that many levels (oriented samples) of the anomalous zone show a considerable scattering of ($\grave{I}_n$) directions. Additionally, in samples B5 and B6 from the Vetovo section, during the demagnetization process the change in ($\grave{I}_n$) direction corresponds to that in ZD (Figs. 1 and 4). For example, in sample B6, the direction of magnetization after demagnetization makes an angle close to 180° with the initial direction, whereas in sample B5, this angle considerably differs from 180° in terms of both ZD and demagnetization results. Two peculiar features were shown in the samples from B6: the first is a small scatter in ($\grave{I}_n$) directions after demagnetization at T = 310°C; the second is that demagnetization virtually did not alter the results obtained at T = 200°C. The directions of magnetization in the samples from the sections above the anomalous zone changed only slightly after demagnetization at T = 310°C.

DISCUSSION OF RESULTS

Thus, the three sections revealed in the lowermost horizons of the buried soil and loess a noticeable decline in factor Q after demagnetization at T = 200°C and considerable changes in the direction of magnetization (even to a reversed direction) after demagnetization at 310°C for 1.5 hours. The relative decline in factor Q in lithologically and magnetomineralogically similar sediments was usually referred to a decrease in

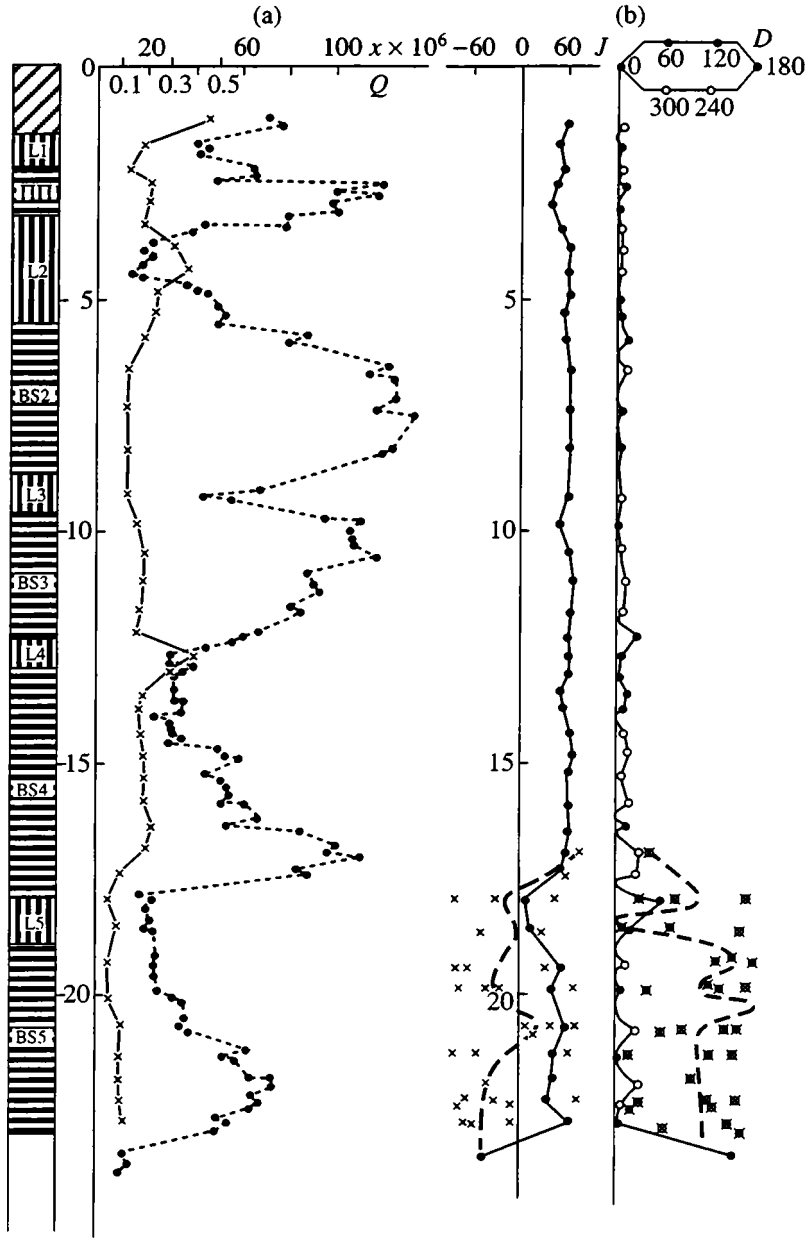

Fig. 2. Lithological, magnetic, and paleomagnetic characteristics of the Kaolinovo section. (See Legend to Fig. 1).

the value or a change in the direction of the ancient geomagnetic field. The compatibility of results obtained after demagnetization at 310°C and from the analysis of the Zijderfeld diagrams allows us to conclude that heating samples in a nonmagnetic environment for 1.5 hours at a temperature of 310°C is sufficient for acquiring the direction of stable magnetization. This conclusion means that the changes in the direction of magnetization with depth should be attributed to changes in the ancient MFE; the latter are probably complicated, as we describe below, by the peculiarities in the formation of the magnetization of the studied deposits.

At first glance, the distribution of the directions of ($\grave{I}_n$) vectors obtained in the lower part of the sections may represent a record of either the Bruhnes event or the Matuyama–Bruhnes transition zone (M–B). We do not rule out this interpretation altogether, but consider it unlikely for two reasons. First, the duration of magnetic and sedimentological events is different. In fact, the maximal duration of the Bruhnes excursion is estimated to be about 10 k.y. (Pospelova, 1989), and that of the M–B inversion, no more than 25 k.y. (Gurarii, 1988). In our case, however, the anomalous zone takes up at least an entire loess-soil horizon, and its age, as implied by its position in the section and the

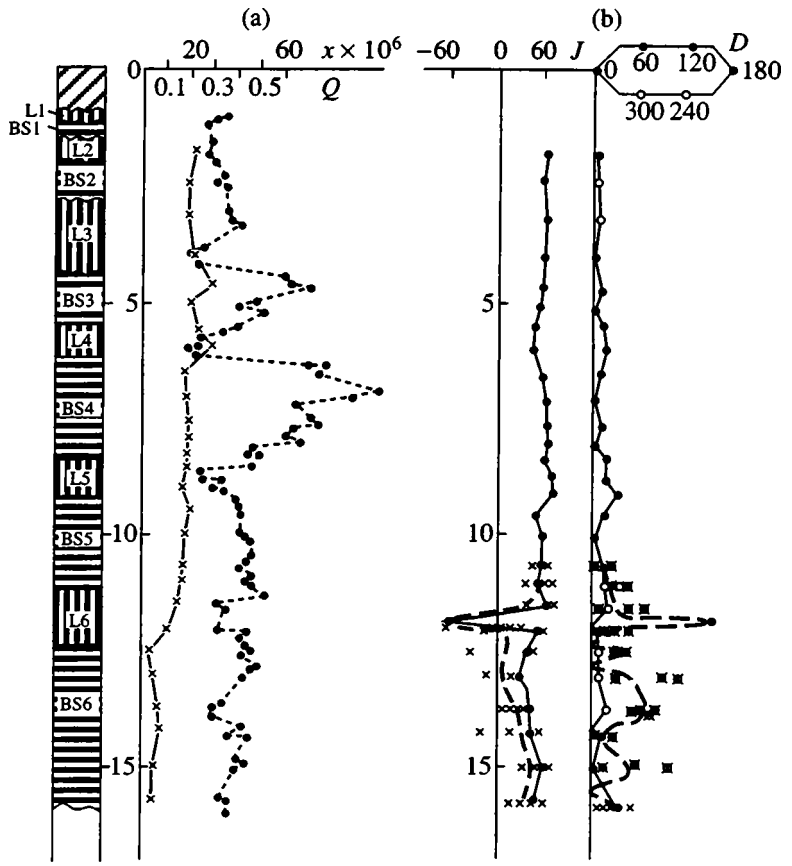

Fig. 3. Lithological, magnetic, and paleomagnetic characteristics of the Durankulak section. (See Legend to Fig. 1).

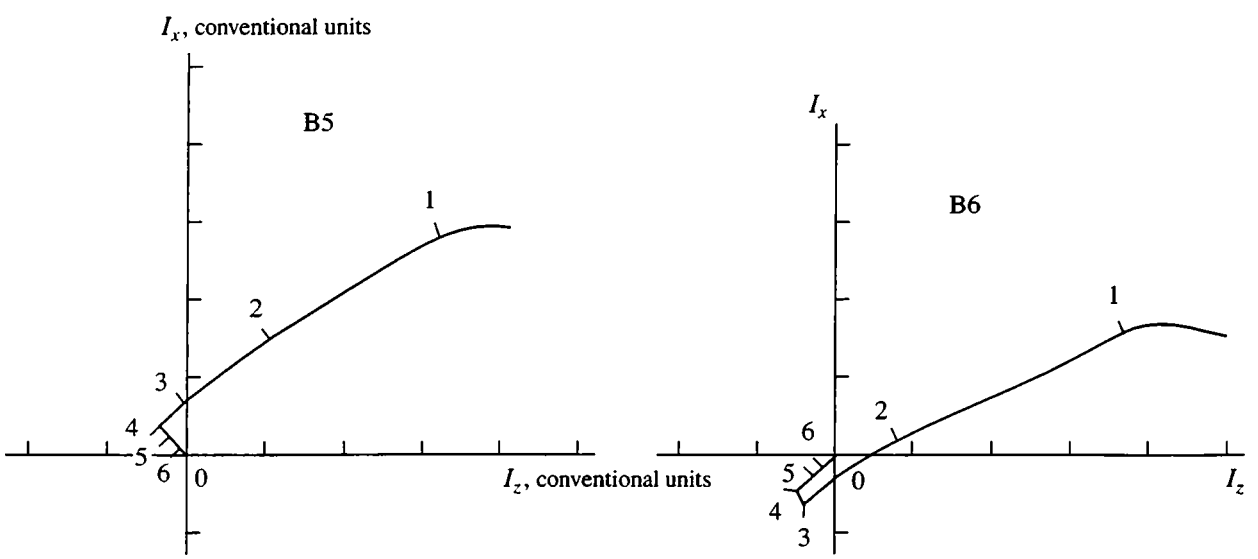

Fig. 4. Variations of components of natural remanent magnetization of samples during heating. 1 – 100; 2 – 200; 3 – 300; 4 – 400; 5 – 500; 6 – 600°C.

type of soil formation, is no younger than the early Pleistocene. This dating is confirmed by the preliminary evidence of a spore-pollen analysis for the Vetovo section obtained by Bolikhovskaya, who considers it possible to attribute the palynoflora of the fourth loess horizon to the Muchkap Interglacial of the early Pleistocene, and the sediments of the fifth loess horizon, to the early Quaternary. The loess-soil complex of northern Bulgaria correlates with the well-studied loess complex of Ukraine and Moldavia (Veklich, 1982) by its structural and magnetic properties and the climatic conditions of its deposition. Veklich has estimated the duration of formation of only one loess-soil horizon of the lower Pleistocene to be no less than 200 k.y. On the other hand, the average rate of loess accumulation in the Pleistocene in much thicker sections in China is estimated by Heller and Tungsheng (1984) to be 7.5 cm/k.y. Assuming even this estimate for the Bulgarian loess, which we believe exaggerated the time of formation of the anomalous loess-soil horizon at Kaolinovo, for example, would be more than 60 k.y., which is incompatible with the duration of the event and is much longer than the inversion period. A second objection is related to the first one in that the Bruhnes epoch, as is universally known, contains up to 12 more or less definitely fixed excursions (Pospelova, 1989). They are rarely observed in continental deposits, primarily owing to the short duration of events and incomplete geological record. When sampling at 50-cm intervals in the upper part of the sections, we could not find a single anomalously magnetized level except the "lightning" anomaly at Vetovo. It seems hardly possible, therefore, that in the comparably thick horizons, constituting 20% of each of the three sections, the magnetization anomaly is associated with an excursion. This means that the magnetization anomaly, which is also characterized by a distinctly reduced Q value, was caused by a more global and continuous MFE variation than the Bruhnes events or the Matuyama–Bruhnes inversion of the geomagnetic field. Therefore, we suggest that the lower loess-soil horizons of the studied sections are coeval and carry a MFE record corresponding not only to the M–B inversion, but to the terminal Matuyama epoch as well. Let us recall that this estimation is in agreement with the palynological data of Bolikhovskaya. The paleomagnetic record of sections at Vetovo and Kaolinovo are, on the whole, in accordance with this conclusion. In the lower part of the loess-soil horizons, the $(\grave{I}_n)$ direction is close to the reversed one, whereas upsection, a certain transition zone can be identified, which is probably still too extended (Figs. 1 and 2). The distribution of $(\grave{I}_n)$ directions in the anomalous zone of the Durankulak outcrop is different; in this locality, the deviation of the direction of magnetization from normal is considerably smaller in the soil layer than in the first two sections (Fig. 3). These discrepancies can be accounted for by the discrete sampling and incomplete geological record. There is yet another circumstance that can essentially affect the records of MFE variations in the loess-soil complex, i.e., the mechanism of formation of magnetization. The problem is complicated and crucial, and though studied by many authors, it remains undeveloped (Veklich, 1982; Tret'yak, 1983; Virina and Faustov, 1973; Heller and Tungsheng, 1984; Hus and Geeraerts, 1986). We shall discuss only two aspects of the problem touched upon in these publications.

The role of bioturbation in the formation of soil has been well-known since the earliest studies. The mechanical mixing of soil naturally affects the orientation of magnetic particles within it, and presumably may result in chaotic directions of $(\grave{I}_n)$ vectors of the magnetic aggregates and a reduction in the resulting magnetization of the soil layer to virtually zero. The results of studying magnetized sediments that fill the burrows, however, testify otherwise, as we know from personal communications from colleagues and as has been proved by our own experience. For example, the paleomagnetic study of the loess-soil sections in the Ob' River area showed that the magnetization of samples collected from a burrow situated in a reversely magnetized loess horizon was normal, i.e., the same as that of the buried soil overlying the loess. Consequently, this means either that the mixing of soil does not considerably disturb the directions of magnetization, or that the process of soil formation includes an additional and more essential factor inducing the soil aggregates to be magnetized in accordance with the direction of the existing MFE. This factor is probably associated with a higher χ value, which is observed in the horizons of steppe and forest-steppe automorphous soils and indicates the prevalence of a highly magnetic mineral, usually magnetite, in them. This kind of magnetite is fine-grained and authigenic (Virina and Faustov, 1973; Maher and Taylor, 1988). The mechanism of its formation is obscure, and it can either be produced by bacteria (Fassbinder *et al.*, 1990) or result from the oxidation of dissolved Fe^{2+} (Maher and Taylor, 1988) or hematite reduction (Virina and Faustov, 1973). If mechanical mixing or the formation of new magnetic minerals develop after the burial of the soil layer, then the magnetic records will lag behind the time of the beginning of the formation of the horizon overlying the soil. In this case, the soil under discussion can be magnetized in the same geomagnetic field as the depositing overlying horizon, for example, loess. Veklich (1982, p. 168) suggests another scenario, according to which the underlying loess horizon is magnetized in the MFE that corresponds to the time of formation of the overlying soil owing to the "vertical migration of ferromagnetites" from the soil formed above. It is particularly important to determine the time delay in magnetic records, as it is essential for special studies of changes in MFE. Heller and Tungsheng (1984) tried to estimate the time of formation of diagenetic hematite, which, as they believe, is responsible for the natural remanent magnetization of the loess complex in China. In their

opinion, the presence of the Jaramillo subchron in the paleomagnetic record indicated that the period of formation of authigenic hematite was an order shorter than that of the subchron, i.e., about 10 k.y.

Our references to the works of Veklich (1982) and Heller and Tungcheng (1984) bear no criticism of their proposals; we only wish to show that the formation of secondary magnetization in a loess-soil complex with a delay of 10 k.y. and longer is accepted by some of researchers without any doubts. Consequently, the above-mentioned processes result in that the initial magnetization in the loess as a soil-forming substance or in the soil would be entirely or partially altered by the "secondary" magnetization, the nature of which can be at the same time chemical, bacterial, and mechanical. The quotation marks around "secondary" are used to indicate that bioturbation, bacterial activity, etc., are integral factors in the formation of a soil layer; without them, this process would be defective, and, therefore, the "secondary" magnetization is, to a certain extent, the initial one. If the formation of the upper part of the discussed loess-soil horizon proceeds during the MFE polarity change, then, due to the magnetic record, delay in the initial and secondary magnetizations could reflect fields with different orientations. In this case, the resulting $(\grave{I}_n)$ will have various directions because of practical noncollinearity of the normal and reversed geomagnetic fields. Thus, the thickness of the transition M–B zone with chaotically directed magnetization can be increased, which is, perhaps, the case in the sections at Vetovo and Kaolinovo. The scatter of $(\grave{I}_n)$ directions in the samples collected from the same level could be caused by the lateral inhomogeneity of the loess-soil layer. This inhomogeneity is apparently confirmed by the significant differences in χ values, occasionally reaching 50%, in samples recovered from one piece of rock. This implies that the more the loess source rock is altered by soil formation processes, the higher its secondary magnetization is.

The stability of the "secondary" magnetization, discussed above, is perhaps comparable with that of the initial magnetization. Therefore, a complete removal of secondary magnetization is impossible without essential distortion of the initial one. As a result, after demagnetization, the density of $(\grave{I}_n)$ directions in normally magnetized samples will be considerably greater than in reversely magnetized samples in which the initial and secondary magnetizations are differently oriented. We cannot accept that this evidence is sufficient for stating that "the obtained data is unreliable," as could be deduced from the second conclusion in the paper by Faustov and Virina (1989). The interpretation of these results is a quite different matter. In general, the practical application of the method of equal density in $(\grave{I}_n)$ directions for estimating the reliability of paleomagnetic results is not as evident as what follows from the studies of Heller and Fungsheng (1984) and Faustov and Virina (1989). In this case, due attention should be given to the fact that the formation of the normal and reversed magnetizations occurred during different time periods. Therefore, the initial magnetization will at least have different ageing characteristics, and the secondary magnetizations will have different values and stability. Also, we can hardly expect that the formation of the normally and reversely magnetized rocks occurred under strictly identical (climatic and other) conditions, which should likewise affect the $(\grave{I}_n)$ density parameters. Consequently, there seems to be no point in trying to achieve the same parameters for normally and reversely magnetized rocks.

The distribution of $(\grave{I}_n)$ directions in the anomalous zone of the Durankulak section, which differs from that in the Vetovo and Kaolinovo sections, can be a result, as we believe, of geomorphologic and other accessory differences in the conditions under which the deposits were formed in these sections. The distribution of χ values in the BS and L in the Durankulak section is somewhat different from that in the other two sections, probably owing to the possible effect of these differences on magnetic properties.

At this stage of research, the variants we suggested may indicate the way to a qualitative explanation (along with other possibilities) for the discrepancies in paleomagnetic data. More concrete representations elucidating the data obtained from these three sections require further extensive studies of the different types of soils, both in the laboratory and in the field.

We shall now make a few comments on the differences between our paleomagnetic results and those of Wiegank (1990). The absence of excursions in our material is due to irregular sampling. Wiegank referred the entire section he studied to the Bruhnes epoch; however, this conclusion is open to questions. In fact, the oldest event, which he identified at the base of the sixth BS and in the seventh L, is, strictly speaking, not an excursion, because the seventh L lies at the base of the section, and there are no data for the underlying strata. For this reason, according to paleomagnetic results alone, this anomalous zone can be either an excursion or a transition zone from the epoch of reversed polarity. Moreover, the data supplied by Wiegank (1990) indicate, in our opinion, that this transition zone should be extended by including in it the entire sixth BS and the lower part of the sixth L. This interpretation practically correlates with our results, which could have been, perhaps, further substantiated if Wiegank had used thermal demagnetization and not demagnetization in an alternating magnetic field (h). In their paper, Heller and Tungsheng (1984) described results obtained by T-demagnetization that were better than those obtained by h-demagnetization of loess deposits. The interpretation of paleomagnetic data obtained by Wiegank was based on the stratigraphic conception that is now prevalent for northern Bulgaria, according to which, the fifth BS is considered to be the time equivalent of the

Mindel-Riss Interglacial, whereas loess accumulation started in the Pleistocene (Wiegank, 1990; Kalcheva *et al.*, 1989). Thus, our paleomagnetic results contradict this stratigraphic conception and correspond better to the previously mentioned paleomagnetic data on the loess complexes of central and eastern Europe.

According to our experience (Bol'shakov, 1984), the application of χ values for the correlation of sections or separate horizons is not an easy task and demands that many factors of deposition be considered. Nonetheless, the problem is sometimes solved, particularly in small regions with similar conditions of sedimentation. There is a striking likeness in the changes of χ values in the Vetovo and Kaolinovo sections. The variations of χ in the Kaolinovo section confirm our supposition that it lacks the fifth loess of the Vetovo section, whereas the visually single BS 4 of Kaolinovo can be divided according to χ behavior into two buried soils. The upper one, above the 15-m mark, has relatively lower χ values and corresponds to BS 4 of Vetovo; the lower half of the two-fold soil with a higher χ value corresponds to BS 5 of Vetovo (Figs. 1 and 2). If we could use this correlation by χ for the Durankulak section, we should then arrive at the conclusion, with due allowance for the correlation between the lower horizons in paleomagnetic data and Q factor, that the Durankulak section between BS 5 and L 6 does not have a loess-soil horizon with a higher χ value of buried soil than that in the overlying and underlying soil horizons, like BS 5 of the Vetovo section. However, taking into account the above-mentioned supposition about the possible differences in conditions of sedimentation in the Durankulak section from the other studied sections, this conclusion lacks sufficient substantiation.

CONCLUSION

The application of methods used in the study of rock magnetism and paleomagnetism to solve various geological and geographical problems, as well as other methods of research, is most successful in a complex study of phenomena. To our regret, we could not obtain any other supplementary data except the reconnaissance results of a spore-pollen analysis. The following conclusions are derived on the basis of magnetic and paleomagnetic studies:

(1) The lowermost loess-soil horizons of the Vetovo, Kaolinovo, and Durankulak sections in northern Bulgaria show a good correlation of magnetic and paleomagnetic data. A more detailed correlation is possible from magnetic data for sections with similar conditions of deposition.

(2) In northern Bulgaria, the loess formation should be dated from the Eopleistocene and not from the Pleistocene, in contrast to the universally accepted opinion.

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